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Deformation, ground subsidence, and slope movements along the Salt River Escarpment in Wood Buffalo National Park

by

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A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Deformation, ground subsidence, and slope movements along the Salt River Escarpment in Wood Buffalo National Park submitted by Po Chow TSUI in partial fulfilment of the requirements for the degree of Master of Science.

ABSTRACT

An escarpment formed by the carbonate Keg River and evaporitic Chinchaga Formations of the Middle Devonian age extends southwards from the Slave River 30 km west of Fort Smith. Hydration of the anhydrite and solution of the gypsum in the Chinchaga Formation underlying the escarpment is causing the overlying rock to deform by folding and faulting.

Field examination of the contact between the Keg River and Chinchaga Formations and construction of the structure contour map on top of the Chinchaga Formation reveal:

1. The area near south Salt River Bridge has been deformed by hydration and expansion of anhydrite, solution subsidence, and normal faulting.
2. Toppling movements occur along the escarpment. It is believed that springs undercutting (causing basal erosion), and frost action on the orthogonally jointed Chinchaga Formation are the main factors responsible for the toppling movement.
3. The area is a gypsum karst area with many distinct karst features such as solution and collapse dolines, anticlinal and homoclinal caves, pocket valleys, half-blind valleys and springs.
4. Ground movement hazards exist in the area. According to the rate of ground movement and its destructive power to

structures and danger to human beings, the study area can be divided into low and high hazardous zones.

5. The escarpment is believed to be the cliffs of Glacial Lake McConnell which appeared at the end of the last glaciation. The last glaciers have also left many beaches along the escarpment. The beaches are the source area for the sand dunes that covered about 30 % of the area.

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Shawinigan Stanley Ltd. for allowing him to examine the cores drilled near Slave River. Fort Smith Flight Service Station provided daily meteorological information for the Altimeter Survey. Dr. J. Toth commented on the groundwater section and his help is acknowledged. The author thanks Mr. Dan Young and Mr. Bim Walters for helpful dicussions. Miss Rita F.N. Chow did the typing of the Appendixes and Tables, Miss Ellie Westbury proof read the manuscripts, their help is greatly appreciated.

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I. INTRODUCTION

A. Purpose of Investigation

1. Little work has been done on the solution and deformation of the carbonate Keg River and gypsiferous Chinchaga Formations which crop out in northeastern Alberta. The study area also consists of gypsum karst and a salt flat which are remarkable and rare features. The investigation may provide a significant contribution to the understanding of the karst processes, the formation of the salt plain and the deformation of the Keg River and Chinchaga Formations.
2. Escarpments which are composed of horizontally bedded rock with orthogonal joints are common. Their retreat has been described and interpreted (Evans, 1981). The study of the escarpment in the area should provide more information on the failure mechanisms typical of these horizontally bedded cliffs under basal weathering and erosion.
3. No study of ground hazards has been carried out in Wood Buffalo National Park. The study area is within the Park and distinct ground hazards are indicated by many fresh solution and collapse dolines, and the toppling failures along the escarpment.

   Recently, several violent "explosions" and uplifts have occurred in Texas which are believed to be due to hydration of anhydrite to gypsum (Brune, 1965);

moreover, solution-subsidence troughs with width and length of 0.1 - 1.0 mile (0.16 - 1.61 km) and 0.5 - 10.0 miles (0.80 - 16.1 km) respectively, have been developed in the gypsiferous Castile Formation (Upper Permian) of the Gypsum Plain in west Texas and southeastern New Mexico (Olive, 1957). The possible occurrence of ground "explosions", slope movement, and rapid to slow ground subsidence in the study area would endanger human lives and structures within the Park. The study of the formation and distribution of these hazards would help to locate the ground hazardous zones within the Park; and furthermore, it can provide valuable information for developing the Park in future, for example; the locations of campsites, highways, fire towers and resort centres.

4. The glacial geology of the area is poorly documented. The investigation should supply knowledge about the effects of the last glaciation on the soluble strata in the area.

B. Location and Accessibility of the Study Area

The study area is located along the northern edge of Wood Buffalo National Park as shown in Figure 1. It extends from latitude 59° 40′ and longitude 111° 54′ - 112° 11′ to latitude 60° 05′ and longitude 112° 25′ - 112° 47′, and extends across the Alberta-Northwest Territories border. The study covers an area of approximately 1185 square kilometres.

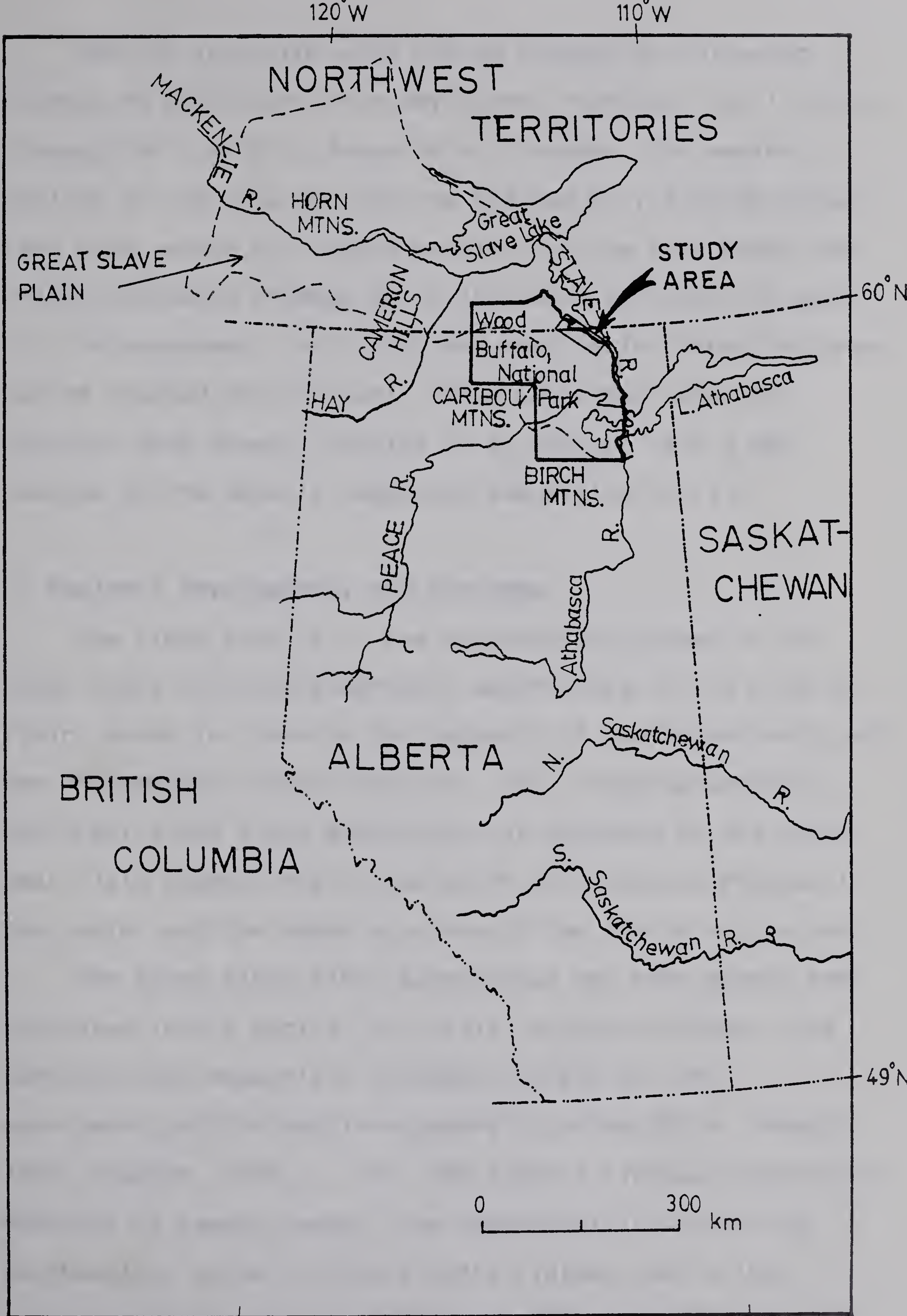


FIGURE 1. LOCATION MAP

Most of the study area can be reached by following Highway #5 which runs from Hay River, Northwest Territories, through Fort Smith to Peace Point. However, the western portion of the area can only be reached by following Parson Lake Road, which is roughly parallel to the Salt River and which intersects Highway #5 in the north and south (Figure 2). The escarpment, salt flat and most of the karst features can be reached only on foot. These walks may range in duration from several minutes to as long as half a day because of the densely vegetated and pitted terrain.

C. Regional Physiography and Drainage

The study area is in the southeastern corner of the Great Slave Plain physiographic subprovince of the Interior Plain, which is close to the boundary of the Borderlands and the Precambrian Shield (Bostock, 1967). Physiographically, the Great Slave Plain subprovince is bordered by the Great Bear Plain subprovince to the north, the Alberta Plateau to the south, and the Kazan province of the Shield to the east.

The Great Slave Plain subprovince has been eroded and glaciated into a gentle, low relief surface; however, the carbonate and evaporitic Paleozoic strata do form escarpments with elevations generally below 300 m (Camsell, 1902; Douglas, 1976, p. 19). The Alberta Plateau subprovince consists of several mesas. The Cameron Hills are on the northwestern corner of the Alberta Plateau; while the Caribou Mountains and the Birch Mountains are situated in

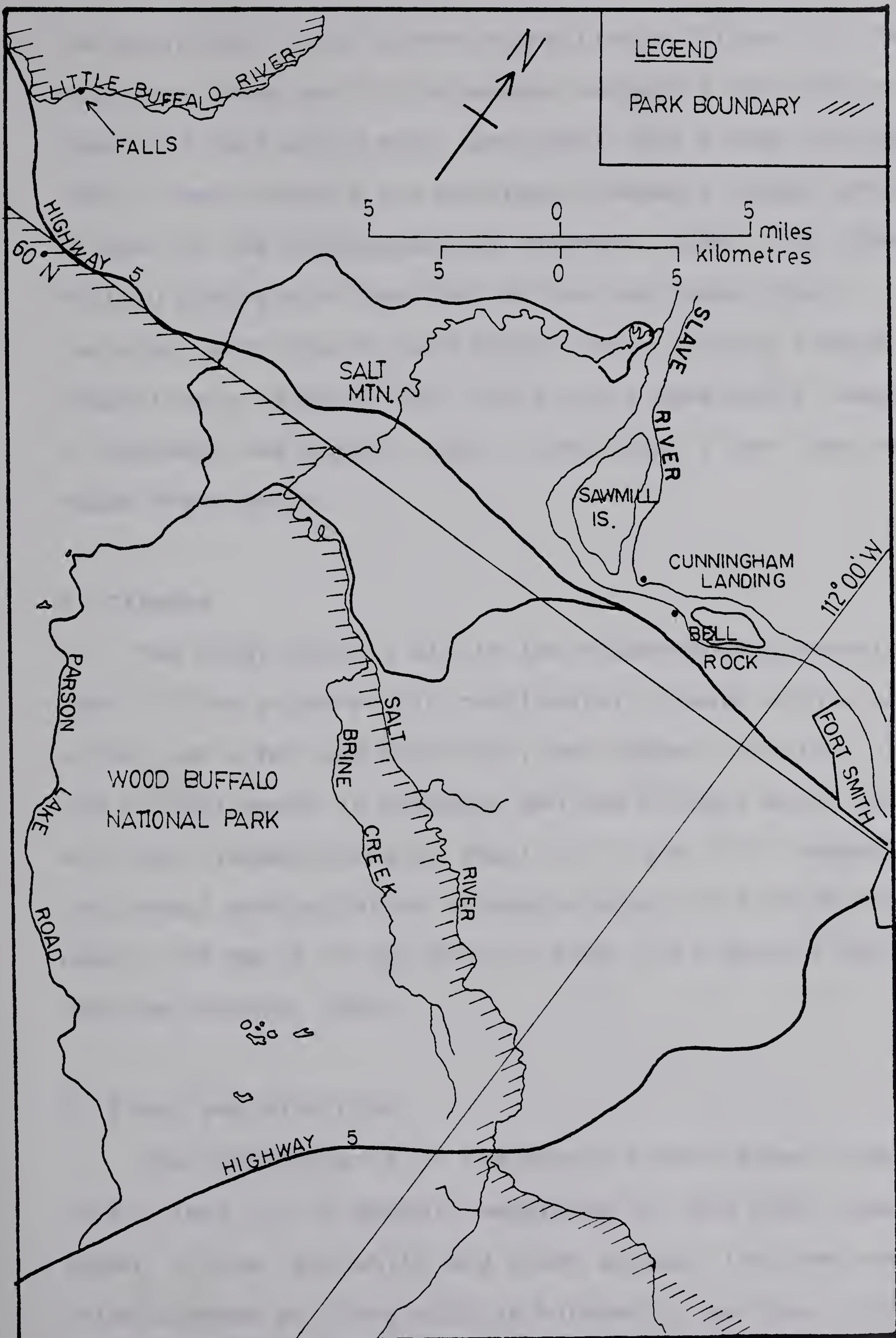


FIGURE 2. SIMPLIFIED GEOGRAPHY — OF THE STUDY AREA

the north and in the centre respectively (Figure 1). These mountains, composed of Cretaceous sediments overlain by layers of Tertiary gravel, are about 1000 m high (Holland, 1964). They probably are erosional remnants formed after the retreat of the Cretaceous sea (Bayrock, 1976). The Alberta Plateau consists of the Fort Nelson and Peace River Lowlands, with the Hay and Peace Rivers running through them respectively. Both of the rivers carry substantial amounts of sediment and deposit them in the Great Slave Lake and the Peace River Delta.

D. Climate

The study area is within the discontinuous permafrost zone. It has a sub-arctic continental climate with a severe winter and a hot and relatively wet summer (Fremlin, 1974). The coldest month is December and the hottest month is July with mean temperatures of about -27 C and 16 C respectively. The annual precipitation is approximately 315 mm of which nearly 109 mm is in the form of snow (Fort Smith Flight Service Station, 1981).

E. Plant and Wild Life

The study area is in the Boreal Forest zone (Fremlin, 1974). Thus, it is densely vegetated by jack pine, tamarack, aspen, willow, and white and black spruce. The open area is often covered by flora such as bluebells, yarrows, wild rye and certain species of weeds (Camsell, 1917). Numerous

marshes, swamps and muskegs are found along the base of the escarpment and on the salt flat. Wood buffalo (bison), black bear, caribou, moose, rabbit, squirrel, beaver, fox and muskrat are also common. Birds in the area include grouse, raven, owl, eagle and whooping cranes. Waterfowl are often seen wandering about marshes, rivers and water-filled dolines.

F. Previous Work

Camsell (1902, 1917) examined the karst landforms southwest of Fort Smith and the Paleozoic gypsum outcrops along the Slave, Salt and Peace Rivers. He was followed by Cameron (1918, 1922a) who described the geological investigation around Great Slave Lake.

The geological work was continued by Govett (1961) and Norris (1963, 1965) who investigated the stratigraphy of Middle Devonian and older Paleozoic rocks around the south Great Slave Lake region and the northeastern portion of Alberta. The glacial features and Glacial Lake McConnell were described by Craig (1965). Belyea (1967), Craig *et al.* (1967), and McCamis *et al.* (1967) interpreted the Devonian facies relationships of northern Alberta and Great Slave Lake region. The Middle Devonian tectonic history from northern Alberta to southern District of Mackenzie has been summarized by Belyea (1971).

In the early 1970s, the surficial geology of the Alberta portion of Wood Buffalo National Park was studied by

the Alberta Research Council (Bayrock, 1972). Subsequently, Bayrock and Reimchen (1976) and Airphoto Analysis Associates (1978) investigated the surficial deposits within the Park and the results are presented as maps and reports (Bayrock, 1976). Recently, the hydrogeology of the gypsum karst and the collapse and solution features in or adjacent to the study area have been described and interpreted by Drake (1970), Cruden *et al.* (1982) and Anonymous (1979).

G. Field Work

The aim of the field work is to document and interpret the nature of solution features, joints, anticlinal structures and the escarpment, as well as to construct several detailed geological maps and cross-sections of the study area in order to determine the mode of structural deformation, movement and solution processes.

The best field season is between the end of May and mid-September when the weather is warm and dry enough for outdoor work. Before the field season started, recent black and white aerial photographs, surficial, topographical and geological maps of the area were obtained and interpreted (Airphoto Analysis Associates, 1978; Bayrock, 1972, 1976).

Suspect outcrops, collapse and subsidence features were located and their access routes were plotted on the photographs. The existing topographic maps of the study area are at a scale of 1 : 250,000 or smaller. All these maps are in Imperial units which make the construction of larger

scale topographic maps in metric units difficult. As a result, Imperial units are used in mapping and throughout this thesis with metric units in parentheses. Appendix 1 shows the conversion factors of these units and other map symbols used in this thesis. Aerial photographs at a scale of 1 : 40,000 and 1 : 50,000 cover the entire study area and are available from the National Air Photo Library.

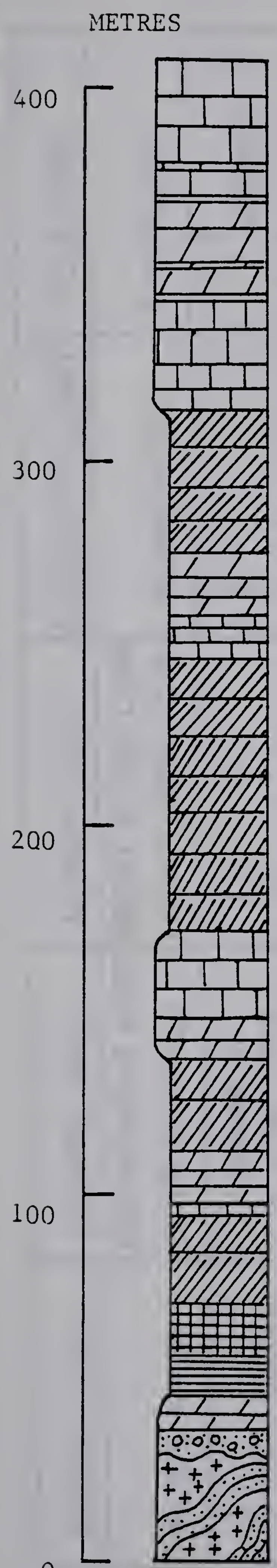
In the field, the writer was accompanied by an assistant. We were equipped with Brunton compasses, barograph, altimeters and surveying tapes. The geologic mapping is chiefly based on the altimeter survey because the area has not been covered by detailed topographic maps. The stable pressure and temperature conditions in the study area enabled the use of single altimeter method in surveying (Hodgson, 1981). Daily meteorological data and topographic heights of several base stations in the area were obtained for the control and corrections of the survey.

II. STRATIGRAPHY

A. Introduction

This thesis concentrates on the Middle Devonian carbonates and evaporites which crop out discontinuously along the Salt and Little Buffalo Rivers, the escarpment west of the Salt River, and within dolines, caves and water-falls behind the escarpment. Studies and correlation of these strata have been performed by Norris (1963, 1965), Craig *et al.* (1967) and McCamis *et al.* (1967). Figure 3 shows a composite generalized stratigraphic section containing all the formations which are exposed within or adjacent to the study area.

Based on lithological observation, the writer has divided the Keg River Formation into Members A, B, and C. The basis for division of the Keg River and Chinchaga Formations and these Members will be discussed in Section C of this Chapter. Figure 4 indicates the correlation of these Members and the nomenclature of various workers. Figure 5 shows a composite stratigraphic section of the Keg River and Chinchaga Formations which are exposed within the study area. The most complete sequence of strata or the type section is exposed along the upper part of Little Buffalo River where it flows over the hard beds of the escarpment in a succession of three falls and extends for about 2 miles (3.2 km) downstream (Norris, 1963, 1965). These three falls are collectively called Little Buffalo River Falls by Norris



AGE	FORMATION	LITHOLOGY
MIDDLE DEVONIAN	SLAVE POINT	DOLOMITIC LIMESTONE, ARGILLACEOUS LIMESTONE, VUGGY, BRECIATED LIMESTONE.
	FORT VERMILION	GYPSUM, SOME DOLOMITIC LIMESTONE, MINOR ANHYDRITE
	NYARLING	GYPSUM, MINOR LIMESTONE, PROBABLY SOME DOLOMITE.
	KEG RIVER	FOSSILIFEROUS, ARGILLACEOUS LIMESTONE, DOLOSTONE, DOLOMITIC LIMESTONE WITH SHALY PARTINGS.
	CHINCHAGA	GYPSUM, ANHYDRITE, DOLOMITE, LIMESTONE, SALT AND MINOR GREEN SHALE.
	FITZGERALD	DOLOSTONE, DOLOMITIC LS.
	LA LOCHE	CONGLOMERATE SANDSTONE, SS.
PRE-CAMBRIAN		GRANITIC GNEISS, PORPHYROBLASTIC FELDSPAR-QUARTZ GRANITES.

FIGURE 3. STRATIGRAPHIC COLUMN OF THE WOOD BUFFALO NATIONAL PARK REGION

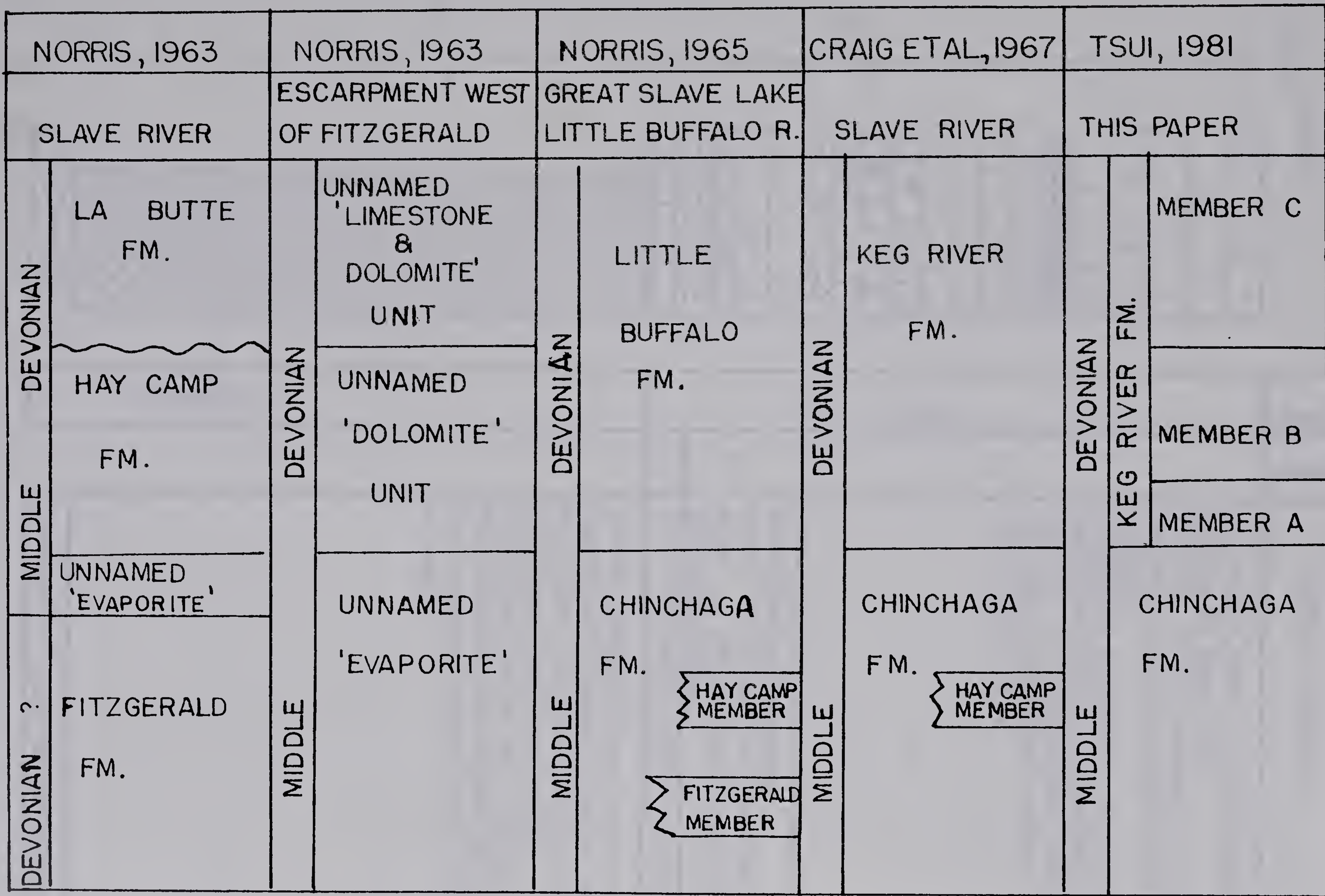


FIGURE 4. STRATIGRAPHIC NOMENCLATURE OF VARIOUS WORKERS AND THIS PAPER

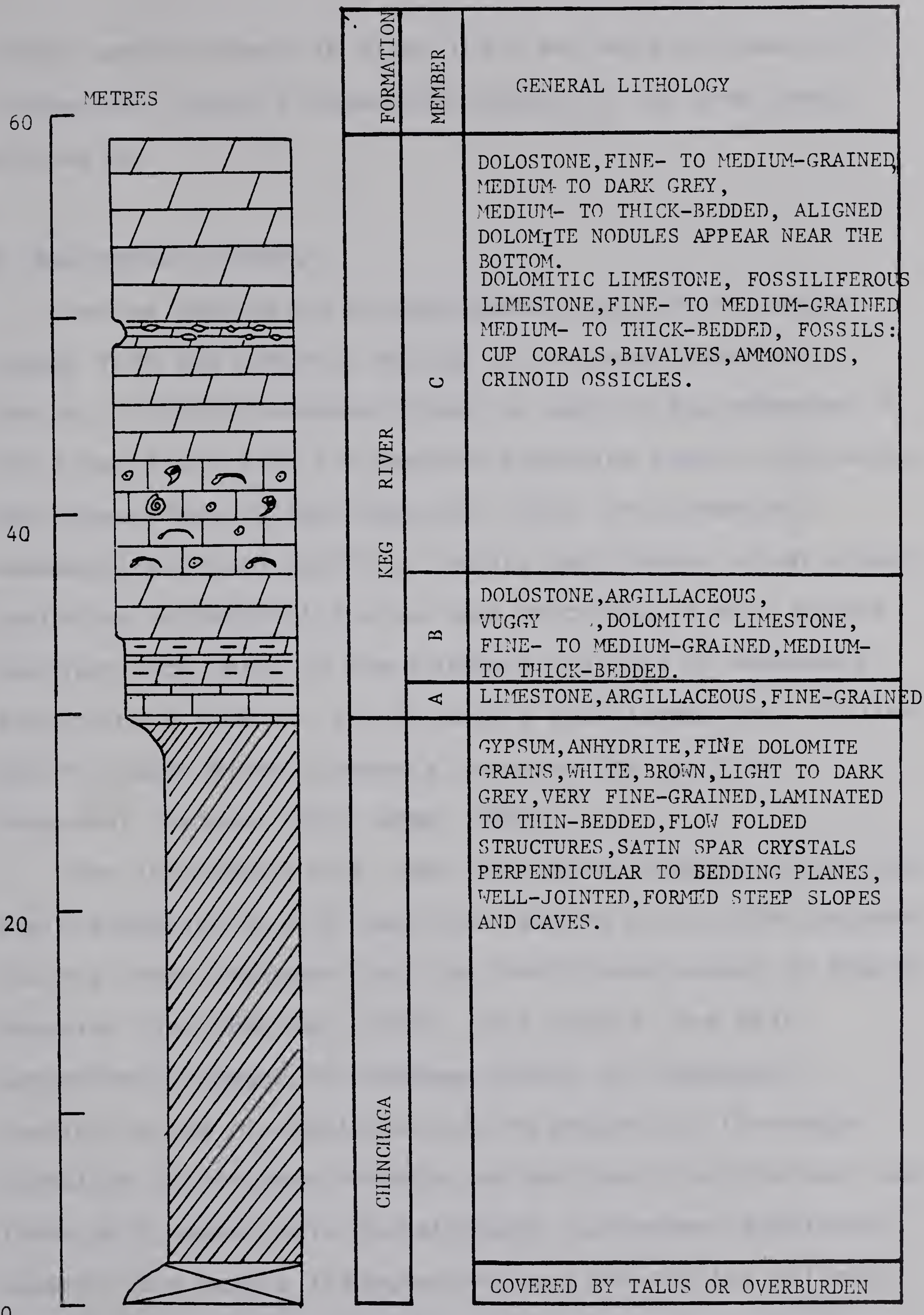


FIGURE 5. STRATIGRAPHY OF THE STUDY AREA

(1963) and are about 39 miles (62.8 km) west-northwest of Fitzgerald. Figure 6 shows the geology of the area under discussion.

B. Geological History

Marine regression and emergence prevailed in western Canada from the Cambrian Period to the Lower Devonian Period. In Middle Devonian time, an epeiric sea appeared in Elk Point Basin with its eastern shoreline roughly following the Slave River Valley (Bayrock, 1976). Pre-Devonian movements produced faulting, horsts and grabens which caused variation in the distribution and thickness of most Middle Devonian rock units in the southern District of Mackenzie and northern Alberta. The movements also formed the Tathlina Uplift, Peace River-Athabasca Arch and the Hay River Embayment (Belyea, 1971; Webb, 1951).

The Tathlina Uplift, the Peace River-Athabasca Arch and the Precambrian Shield restricted marine circulation between the Hay River Embayment and the Cordilleran seaway in Middle Devonian time (Bayrock, 1976). As a result, the salt concentration began to increase within the Embayment, leading to the precipitation of the evaporitic Chinchaga Formation in northern Alberta and Northwest Territories, and thick salt deposits in Saskatchewan. Subsequent subsidence caused rapid marine transgression and diluted the salinity in the Embayment which resulted in a carbonate sedimentary environment. Fringing reefs started to flank the highlands

such as the Peace River Arch (McCamis *et al.*, 1967, p. 444). Subsequently, the carbonate Keg River Formation and other Middle and Upper Devonian strata were deposited.

Deposition did not resume until Cretaceous time when a shallow sea invaded the Great Slave Lake region and deposited clastic materials. Uplift continued in the Tertiary Period and Wood Buffalo National Park was chiefly under pedimentation (Bayrock, 1976, p. 41). Finally, the Wisconsin-Laurentide ice sheet glaciated western Canada several times during the Pleistocene Epoch. Craig(1965) stated that there were at least two glaciations in the region of Mackenzie.

C. Strata Description

Chinchaga Formation

The Chinchaga Formation is estimated to be 91 m thick and consists of gypsum, anhydrite, dolomite, limestone and salt (Norris, 1965; Craig *et al.*, 1967). It overlies the Fitzgerald Formation (though its bottom contact is not exposed) and is overlain conformably by the Keg River Formation. The top of the Chinchaga Formation is defined by the first appearance of a dark grey to black, fine-grained and thin-bedded limestone. This is clearly exposed near the base of the main falls on Little Buffalo River (Appendix 2, Section 1).

Within the study area, the Chinchaga Formation is best exposed downstream of the main falls on Little Buffalo

River, in the Walk-in Cave, in the Dome Collapse Doline, along the banks of the Salt River near south Salt River Bridge, and along the escarpment overlooking the salt flat. The type section of the Chinchaga Formation is exposed immediately below the main falls on Little Buffalo River and extending for about 2 miles (3.2 km) downstream (Norris, 1963, 1965). Figure 7 indicates the location of these outcrops which were examined by the writer and Appendix 2 shows the geologic descriptions of the sections measured in these exposures. The Little Buffalo River section has been described particularly by Norris (1963), and the outcrops along the escarpment were briefly mentioned by Camsell (1917). The stratigraphy of the remaining sections is herein being documented for the first time.

Except the beds exposed around south Salt River Bridge, the outcrops indicate that the Chinchaga Formation is a white, banded light to dark grey, brown to dark brown, laminated to very thin-bedded, coarse- to fine-grained, pure to argillaceous gypsum and anhydrite, with shaly breccia and dolomite inclusions (Appendix 2, Sections 2, 3 and 10). The Chinchaga Formation, observed along Salt River, is a powdery white to light bluish grey, laminated to medium-bedded, fine-grained gypsum and anhydrite. The bedding is undulatory and contorted. The gypsum and anhydrite may display intricate folds and individual nodules distorted against each other; moreover, thin layers of clay are found surrounding these nodules and between beds. These features

are clearly shown in outcrops along Salt River, and also in the gypsum exposed at Peace Point along Peace River which have been described by Halferdahl (1960) (Figure 7; Plates 1 and 2; Appendix 2, Sections 6 and 9). All measured sections show layers of satin spar with crystals about 1 cm long which tend to grow perpendicular to the bedding planes.

The Chinchaga Formation is poor in fossils and hence it is dated as Middle Devonian based chiefly on the Middle Devonian fossils collected from the Keg River Formation above and the Fitzgerald Formation below (Norris, 1965, p. 6).

Keg River Formation

Norris (1963, 1965) and Craig *et al.* (1967) described the Keg River Formation which is exposed along the Slave River in northeastern Alberta and at the water-falls on Little Buffalo River in Northwest Territories. The Keg River Formation is estimated to be approximately 35 m thick. It is described as ranging from a medium brown argillaceous and fossiliferous limestone, to a pale to dark brown, aphanitic to medium-grained, vuggy to gypsiferous dolomite (Norris, 1963, 1965). It overlies conformably the Chinchaga Formation with a sharp contact. The contact is drawn at the boundary between a light to medium grey laminated gypsum and a dark grey to black, fine-grained and thin-bedded limestone (Appendix 2, Section 1). The upper contact with the overlying Nyarling Formation is not exposed.

Plate 1. Intricate folds in gypsum outcrops near south Salt River Bridge.

Plate 2. Gypsum nodules in outcrops near south Salt River Bridge.

Within the study area, the Keg River Formation outcrops within the Walk-in Cave Doline, the Fire Tower Doline, the three falls on Little Buffalo River and along the Salt River near south Salt River Bridge as shown in Figure 7. Sections 1, 4, 5, 7 and 8 in Appendix 2 show the lithologies of these outcrops.

Norris (1963, 1965) examined the carbonate sections exposing along the Slave and Little Buffalo Rivers and divided them into two units: the lower 'Dolomite' and the upper 'Limestone and Dolomite'. However, the writer have divided the Keg River Formation into three Members based on lithological differences observed in the study area. They are Members A, B and C. Members A and B are stratigraphically equivalent to the 'Dolomite' unit and Member C is equivalent to the upper 'Limestone and Dolomite' unit designated by Norris (1963). The type section of these Members is located at the three falls on Little Buffalo River.

Member A

Member A is about 2 m thick and is a dark greenish-grey to black, very fine-grained, laminated to thin-bedded fossiliferous limestone. Near the base of the Member, the bedding is commonly folded with amplitudes and wavelengths of about 3 cm and 12 cm respectively. The lower beds are also locally brecciated. Member A is found at the base of the Keg River Formation that overlies conformably the Chinchaga

Formation with a sharp contact. The bottom of Member A is defined as the boundary between the Keg River and Chinchaga Formations. The top of the Member is distinct and is drawn at the sharp contact with a light brown to grey, very fine-grained, faintly laminated, argillaceous dolostone to dolomitic limestone.

Member B

At the main falls on Little Buffalo River, Member B is about 11.2 m thick and overlies conformably Member A and underlies conformably Member C. The top of Member B is defined by the first appearance of a medium to dark grey, well-jointed, thin- to thick- bedded limestone.

The measured sections show that Member B is a light grey to dark brown, coarse- to fine- grained, thin- to thick-bedded vuggy dolostone. There is a light brown to grey, very fine-grained, faintly laminated, argillaceous dolostone to dolomitic limestone about 2 m thick present at its base.

Member C

The upper contact of Member C with other members (if any) or the Nyarling Formation has not been observed in the study area; however, the sections measured at the Little Buffalo River Falls and the Walk-in Cave Doline indicate that it overlies Member B conformably. Member C is at least 12.5 m thick and can be divided into two distinct units.

The lower unit is about 5.5 m thick and is a medium to dark grey, fine- to medium-grained, thin- to thick-bedded fossiliferous limestone. Generally, the fossils are more abundant and larger in size near the bottom of the unit, and they tend to decrease upwards both in number and size.

Fossils identified include: *Atrypa* sp. cf. *A. arctica* Warren, *Gypidula* sp., gastropods, cup coral, cf. *Michelinoceras* sp., sponge spicules, and crinoid ossicles. In fact, the Keg River Formation is dated as Middle Devonian mainly because of the presence of the *Arctica* zone (Norris, 1965; Craig *et al.*, 1967). The top of the lower unit is defined by the first appearance of a vuggy dolostone to dolomitic limestone.

The upper unit is about 7.2 m thick and is essentially a light grey to brown, fine- to medium-grained, thin- to medium-bedded vuggy dolostone to dolomitic limestone.

D. Correlation of the Keg River Formation

Figure 8 shows that the Keg River Formation exposed in the study area can be correlated based on the three Members introduced by the writer. Figure 7 shows the location of these exposures. The datum is drawn at Elevation 610 ft (186 m) above mean sea level.

The sharp contacts of the Keg River and Chinchaga Formations, and of the Members above indicate that rapid

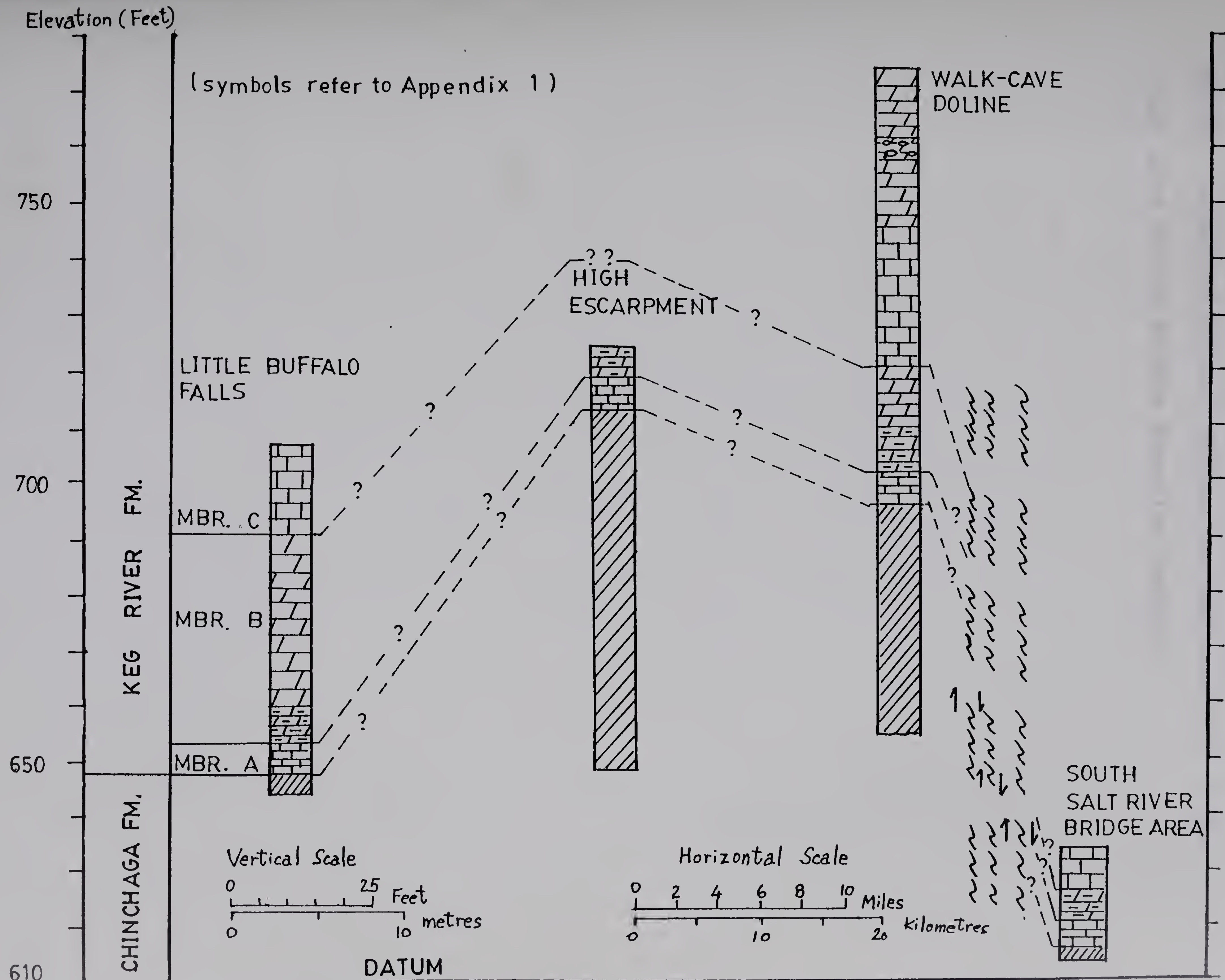


FIGURE 8. STRUCTURAL CROSS-SECTION ON A 610-FOOT DATSUM

marine transgression and regression had occurred in the study area during Middle Devonian Period.

III. GLACIAL GEOLOGY

A. Introduction

The area was glaciated by the Wisconsin-Laurentide ice sheet which advanced from the District of Keewatin and extended westward into the Mackenzie Mountains. The ice was at least 2000 ft (610 m) thick in Wood Buffalo National Park area (Craig, 1965, p.29; Bayrock, 1976). The retreat of this ice sheet back to the Precambrian Shield created a vast lake known as Glacial Lake McConnell (Craig, 1965). The glaciers and the wave action of this Glacial Lake produced features such as wave-cut platforms, cliffs, beaches, off-shore bars and a broad plain of glaciolacustrine deposits.

It is believed that subsequent reworking of the lake sediments (abandoned beaches) by wind has resulted in the formation of sand dunes in the area (Douglas, 1959, p. 7). Due to the dense vegetation growth in the study area at present, the dunes have been well stabilized (Douglas, 1959, p. 7; Bayrock, 1976, p. 26).

B. Glacial Deposits and Glacial Features

Till

A mantle of till is present at and behind the escarpment in the study area. The till is silty and clayey in texture with many subangular to subrounded granitic to gneissic boulders and gravels, and locally derived carbonate and gypsum pebbles. They are bound tightly together by

calcium carbonate cement and usually form a steep wall above the exposed bedrock. For instance, the till layer exposed in the Fire Tower Doline has a slope of about 40 degrees.

Generally, the till is 1 - 12 m thick and has the level to gentle rolling topography characteristic of ground moraine on flat-lying strata (Bayrock, 1976). The till measured along the High Escarpment varied from 3 m to 6.5 m thick, and the till examined in Walk-in Cave Doline and along the banks of Salt River in the vicinity of south Salt River Bridge ranged from 0.15 - 2.4 m thick (Appendix 2, Sections 5, 7 and 10).

Bayrock (1976) stated that the composition of till in Wood Buffalo National Park reflects the composition of the underlying bedrock. The till exposed along the High Escarpment where the Chinchaga Formation outcropped is chiefly made up of gypsum fragments (Figure 7; Appendix 2, Section 10); while farther back from the escarpment where the carbonate Keg River Formation is close to the surface (for instance, in Walk-in Cave Doline), the till is composed of numerous limestone and dolostone pebbles (Figure 7; Appendix 2, Section 5).

Glacial striations were observed at outcrops exposed along the banks of the Salt River about 2.6 km upstream of the intersection of south Salt River Bridge and the River; and on the surface of Parson Lake Road, which is approximately 6.15 km south of the turn-off from Highway #5 going west to Hay River (Figure 9). These striations are

millimetres wide and some run continuously for a few metres.

The glacial striations exposed along Salt River are found on weathered fossiliferous limestone and show an ice movement at about 260 degrees; while those on Parson Lake Road are found on dolostone and indicate an ice flow direction of roughly 222 degrees. They are comparable with the glacial striations observed by Craig (1965) along Hay River which indicate two directions of glacial movement at approximately 230 and 310 degrees.

Craig (1967) noticed that the southwesterly oriented striations are more numerous and better preserved suggesting that they were formed later than the northwesterly striking set (Craig, 1965, p. 28). Similarly, Whittaker(1922, 1923) observed striations on Trout and RedKnife Rivers, which are about 110 Km west of Fort Providence, and stated that the southwesterly set was younger.

On this account, the last ice which glaciated the study area probably moved along a line having an azimuth approximately 222 - 260 degrees.

Glaciolacustrine Deposits - Glacial Lake McConnell

At the end of the last glaciation, a proglacial lake known as Glacial Lake McConnell had formed. It extended from the north side of Great Bear Lake, occupied the whole Great Slave Lake Basin to the lower parts of the valleys of Peace and Athabasca Rivers and part of the west portion of Lake Athabasca Basin (Craig, 1965, p. 18). When the ice-front

continued to retreat northeastward and further isostatic adjustment occurred, the water level of the glacial lake was greatly lowered such that the waters of the three basins, Great Bear Lake, Great Slave Lake and Lake Athabasca, became separated (Cameron, 1922b, p. 351 - 353). At that time, the depression between the Slave River and the escarpment in the study area was covered by the ancient arm of Great Slave Lake (Cameron, 1922a, p. 10, 1922b, p. 351; McConnell, 1890).

The glaciolacustrine deposits are found chiefly on the depression in front of the escarpment. They are mainly silty clay to fine sand, and are derived from sediments carried by meltwater into the glacial lakes which existed during and after the retreat of the continental glacier (Bayrock, 1976, p. 25). The deposition of these glaciolacustrine sediments formed a vast and flat lake basin which became a salt plain after the lake was drained.

The thickness of these deposits has not been measured in the field. Bayrock (1976, p. 25) reported that drilling near Lake Claire, which is about 130 km south of the study area, had revealed over 61 m of glaciolacustrine sediments. Terrain analysis by Bayrock and Reimchen (1976) shows that the glaciolacustrine deposits in most of the study area is over 12 m thick. Drillings reveal that the glaciolacustrine deposits are about 8 - 30 m thick around Fort Smith which is about 21 km east of the study area (Shawinigan Stanley Ltd., 1981).

The shoreline of the former arm of Great Slave Lake can be determined by the location of the abandoned beaches, terraces, lake-cliffs, off-shore islands and the escarpment. Figure 9 illustrates that the abandoned beaches and terraces identified on the aerial photographs only appear between the 600 ft (183 m) and 800 ft (244 m) contours in the study area. The 800 ft (244 m) contour seems to be the highest elevation reached by the ancient arm of Great Slave Lake. Indeed, Craig (1965) has drawn the approximate extent of Glacial Lake McConnell, and claimed that abandoned beaches found at the southern flank of Horn Plateau, which is located west of the north arm of Great Slave Lake (Figure 1), had never exceeded the 800 ft (244 m) contour.

In addition, the steepest portion of the escarpment, for example, the High Escarpment, faces southeast (Figure 9). Bayrock (1976, p. 26) pointed out that the wind mainly blew from the southeast in Wood Buffalo National Park immediately after the last glaciation. This wind would create waves on the proglacial lake which would attack directly that portion of the escarpment facing southeast. Thus, there would be steep cliffs but no beach deposits along southeast-facing portions of the escarpment; on the other hand, the rest of the escarpment would have gentler slopes and well-developed beaches.

Figure 9 shows the distribution of abandoned beaches and the directions of the faces of the escarpment.

C. Dunes

Air photo interpretation indicates about 30 % of the study area is covered by aeolian deposits (Airphoto Analysis Associates, 1978). The deposits are mainly composed of medium to fine sand and vary from 1 m to over 12 m thick. To illustrate, the Fire Tower Doline has exposed 2 m of sand (Appendix 2, Section 4). They are easily identified on air photographs and can be classified into U-shaped dunes, longitudinal dunes and sheet sand (Figure 9).

U-shaped Dunes

U-shaped dunes resemble barchan dunes. However, barchans are typical of desert areas, while U-shaped dunes developed in conflict with surface vegetation, are typical of the semi-arid to humid areas (Odynsky, 1958, p. 57; Gravenor *et al.*, 1978). Thus, based on the semi-arid climate and dense vegetation in the study area, the roughly crescent-shaped hills of sand are classified as U-shaped dunes. Note that U-shaped dunes always have their concave sides and horns facing the wind.

On aerial photographs, it has been identified that some U-shaped dunes have their concave sides and horns facing southeast; while some of them have theirs facing northwest. This indicates that dunes in the area were formed by wind coming from opposite directions at different times, that is, southeast and northwest.

The study of the U-shaped dunes in northeastern Alberta and the meteorological maps of Canada by Odynsky (1958) indicate that the effective wind direction responsible for the formation of the U-shaped dunes in northeastern Alberta at present is from the northwest (Odynsky, 1958, p. 58, figure 3; Gravenor *et al.*, 1978, p. 40). Bayrock (1976, p. 26) investigated the sand dunes in Wood Buffalo National Park and concluded that immediately following the recession of the last glacier and drainage of glacial lakes, the wind blew from southeast, and the present sand dune forming wind direction is roughly from northwest.

In the study area, the U-shaped dunes with horns about 200 - 600 m long and 3 - 45 m thick are common.

Longitudinal Dunes

Longitudinal dunes are sinuous ridges and may reach 5 km long in the study area. Bayrock (1976) reported that longitudinal dunes up to 32 km long existed in the north of the Park.

Sheet Sand

Sheet sand appears between U-shaped and longitudinal dunes and is a few metres thick. Fire Tower Doline is probably located in a sheet sand area.

IV. STRUCTURE

A. Introduction

Three major fabric elements are observed in the study area. They are : 1) joints 2) anticlines and depressions 3) faults.

Field measurement and structural projection indicate that the top of the Chinchaga Formation in the northern portion of the study area has a strike and dip of 146 degrees and 0.17 degrees SW respectively; however, the strike swings south in the southern portion of the area with a gentler dip of 0.14 degrees W (Figure 10). This is comparable to Belyea's (1971) structure contour maps on top of the Chinchaga Formation in southern District of Mackenzie and northern Alberta which show a 155 degrees regional strike and 0.23 degrees SW dip; and, the maps also indicate that the contours swing southward from about 160 degrees at the top of the Chinchaga Formation to about 170 degrees on the Fort Vermilion marker (Belyea, 1971, p. 31, figure 22). This change is interpreted as a change in the direction of transgression of successive Devonian invasions probably associated with regional warping (Belyea, 1971, p. 32). Figure 10 is a structural contour map on top of the Chinchaga Formation of the study area.

The attitudes of the Keg River and Chinchaga Formations and other Paleozoic rocks are controlled by the underlying undulatory Precambrian basement. Shawinigan Stanley Ltd.

(1980; 1981, drawing no. 2001-A0-1265) performed drilling and geophysical studies east of the study area between the Salt and Slave Rivers and showed that the Precambrian surface has a dip of about 0.18 - 0.23 degrees to the southwest. Norris (1965, p. 86) stated that the Precambrian surface has a regional strike of 150 degrees and dip of 0.22 degrees SW in the Great Slave Lake region.

The above observations are in agreement with the regional homoclinal structure under the plains of northeastern Alberta and adjacent territories with dips of 0.22 - 0.27 degrees SW (Law, 1955, p. 1971).

B. Joints

In the study area, two orthogonal joint systems are found in the Chinchaga Formation exposed along the High Escarpment and downstream of Little Buffalo River. System I joints consist of sets striking 54 and 150 degrees which are approximately perpendicular and parallel to the regional strike of the bedding; System II joints trend about 6 and 86 degrees (Figure 11). On the other hand, it seems that there is only one orthogonal joint system in the Keg River Formation that is observed in the Fire Tower and the Walk-in Cave Dolines, and the main falls on Little Buffalo River, with the sets striking 66 and 158 degrees (Figure 11).

The dip of these orthogonal joint systems are roughly normal to the nearly flat-lying strata in the study area. The south Salt River Bridge area is excluded in joint

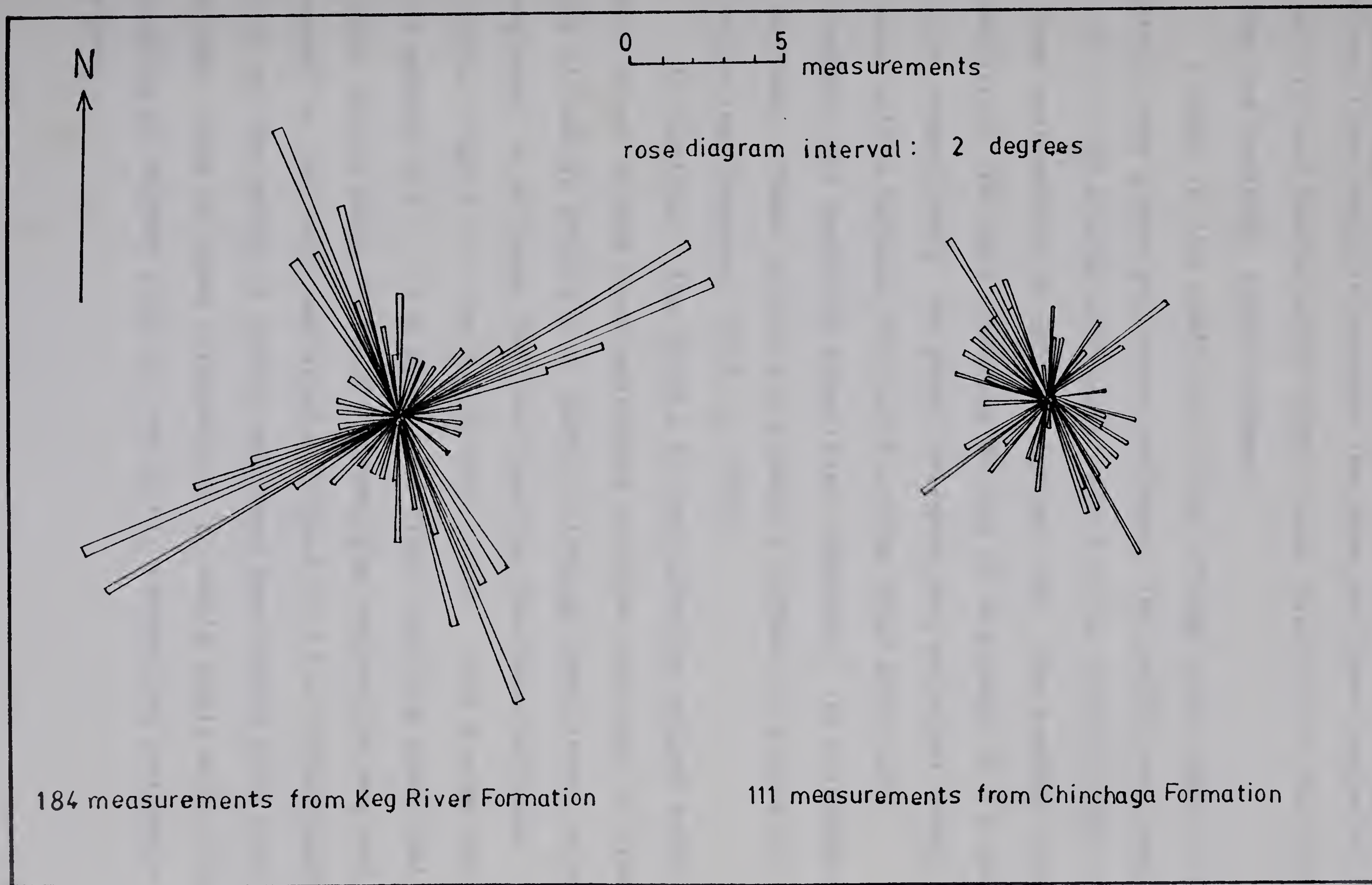


FIGURE 11. DISTRIBUTION OF JOINTS IN THE KEG RIVER & CHINCHAGA FORMATIONS

orientation analysis because the area had been subjected to faulting and doming which might cause rotation of the joints from the regional joint systems.

Except in the south Salt River Bridge area, most of the joints show clean and fresh surfaces. The joints are usually 0.10 to 0.60 m wide with very little or no infills; and from a few centimetres to tens of metres long. The average joint spacing of the sets trending NE and SE are about 6.4 m and 4.4 m respectively. No shear movement has occurred along these joints because the joint planes show sharp and irregular but matching asperities; moreover, no slickensides are found. These joints are probably formed by extension fracturing (Babcock, 1973, p. 1779).

In contrast, the joints in the outcrops around south Salt River Bridge have weathered surfaces and most of them are filled by crystals or debris although some fresh joints are seen in collapse features such as the Dome Collapse Doline (Figure 7). The joints are 0.006 - 0.40 m wide and are spaced 0.3 - 1.2 m apart. Vertical sections show that these joints have 46 to 90 degrees dip and may or may not be normal to the inclined strata. Figure 12 is a stereographic plot of the poles of joints measured near south Salt River Bridge.. Some of them intersect at an acute angle with offset bedding on either side of the joints resembling conjugate shears.

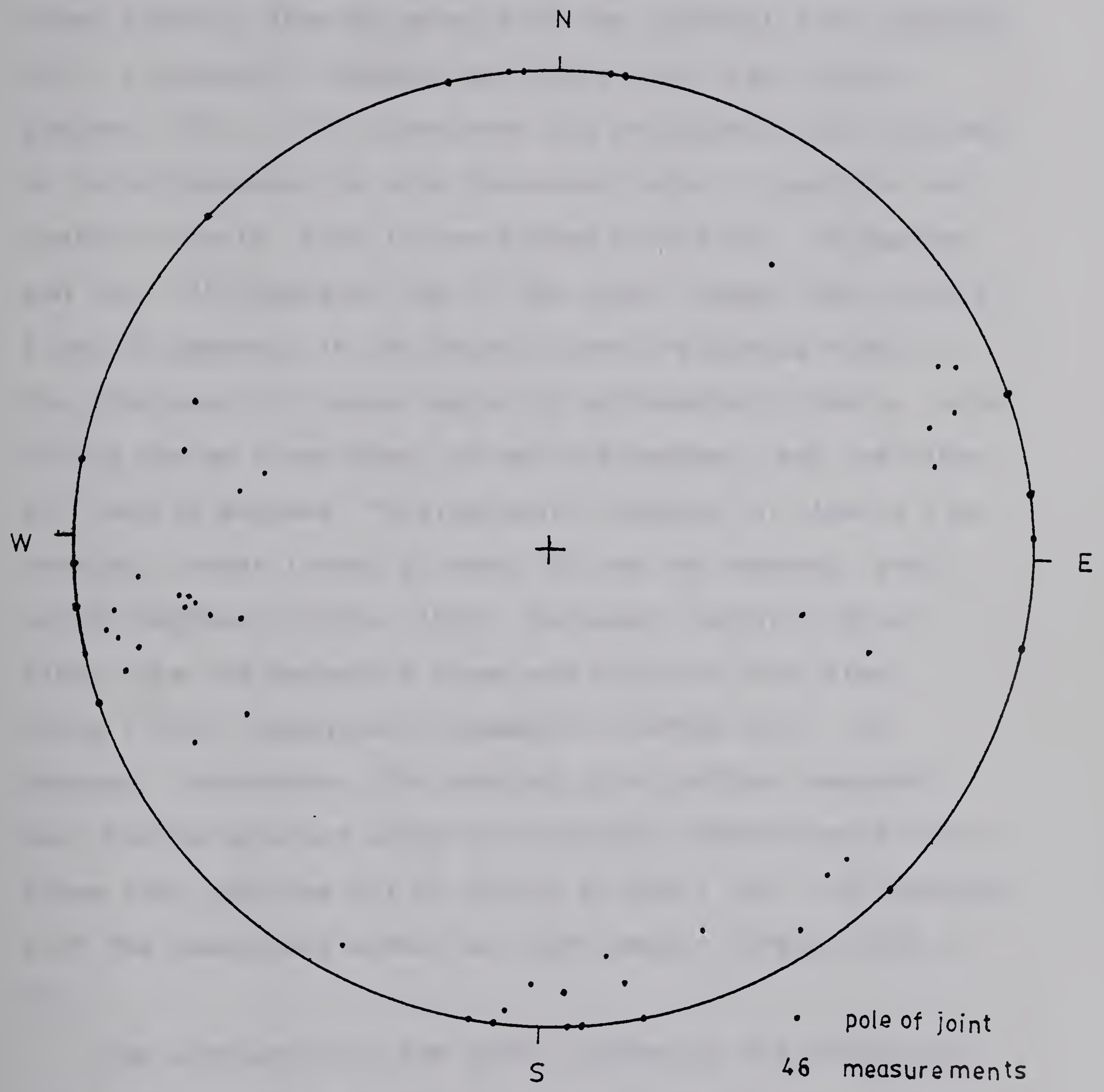


FIGURE 12. : STEREOGRAPHIC PLOT OF THE POLES OF JOINTS MEASURED AT SECTION 7 NEAR SOUTH SALT RIVER BRIDGE

Although the orthogonal joint systems in the Chinchaga and Keg River Formations of the study area do not match each other exactly, they do agree with the regional joint pattern that is present in Alberta and Great Slave Lake region. Babcock (1973, 1974) discovered two orthogonal joint systems in Late Cretaceous to Late Paleocene rocks of southern and central Alberta. Sets in one system strike 55 - 65 degrees and 140 - 155 degrees, and in the other system, sets strike 5 and 95 degrees. In the Devonian and Cretaceous rocks in the Athabasca Oil Sands region in northeastern Alberta, sets in one system trend about 50 and 145 degrees; and the other at 0 and 90 degrees. Physiographic features in Alberta also indicate linear trends at about 54 and 144 degrees, and 0 and 90 degrees (Ozoray, 1972). Moreover, north of Great Slave Lake and Mackenzie River and north of Horn River, Craig (1965) recognized lineaments oriented 140 - 160 degrees. Furthermore, the bedrock joint pattern measured near Kakisa Lake and along the northwest shoreline of Great Slave Lake show one set of joints at about 140 - 160 degrees with the second set almost at right angles (Craig, 1965, p. 27).

The similarity of the joint systems of the study area with the regional jointing in western Canada suggest that they may be formed by the same mechanism. The dominant joint systems in the study area have sets striking about 60 and 154 degrees; which is approximately normal and parallel to the trend of Cordilleran orogenic belt, and the extensional

origin of these joints seems to suggest that they may be related to residual stresses from past tectonic events such as the Laramide compression of the Rocky Mountains (Babcock, 1974, p. 1185; Bell and Gough, 1979).

C. Anticlines and Depressions

Description

Anticlinal structures in the study area only appear near south Salt River Bridge area where they are exposed along the banks of Salt River, and in caves and scarps behind the River. Figure 13 shows the location of these deformed structures.

Commonly, the Keg River Formation has been deformed into gentle folds with an average wavelength of 610 m and an amplitude of 12 m. However, Member B of the Keg River Formation can be deformed into close folds; while the overlying Member C still maintains gentle folds (Plate 3). The terminology used in describing the tightness of the folds in this thesis follows Fleuty (1964, p. 470).

The Chinchaga Formation tends to have folds with a wavelength of 70 m and an amplitude of 17 m (Figure 14 and Plate 4). These gypsiferous rocks had been deformed into dome-shaped structures. This is shown by a doline called here the Dome Collapse Doline near Salt River with beds dipping away from its centre (Figure 15).

Figures 16 and 17 also indicate an uplift with a central depression having a diameter of about 1280 ft (390

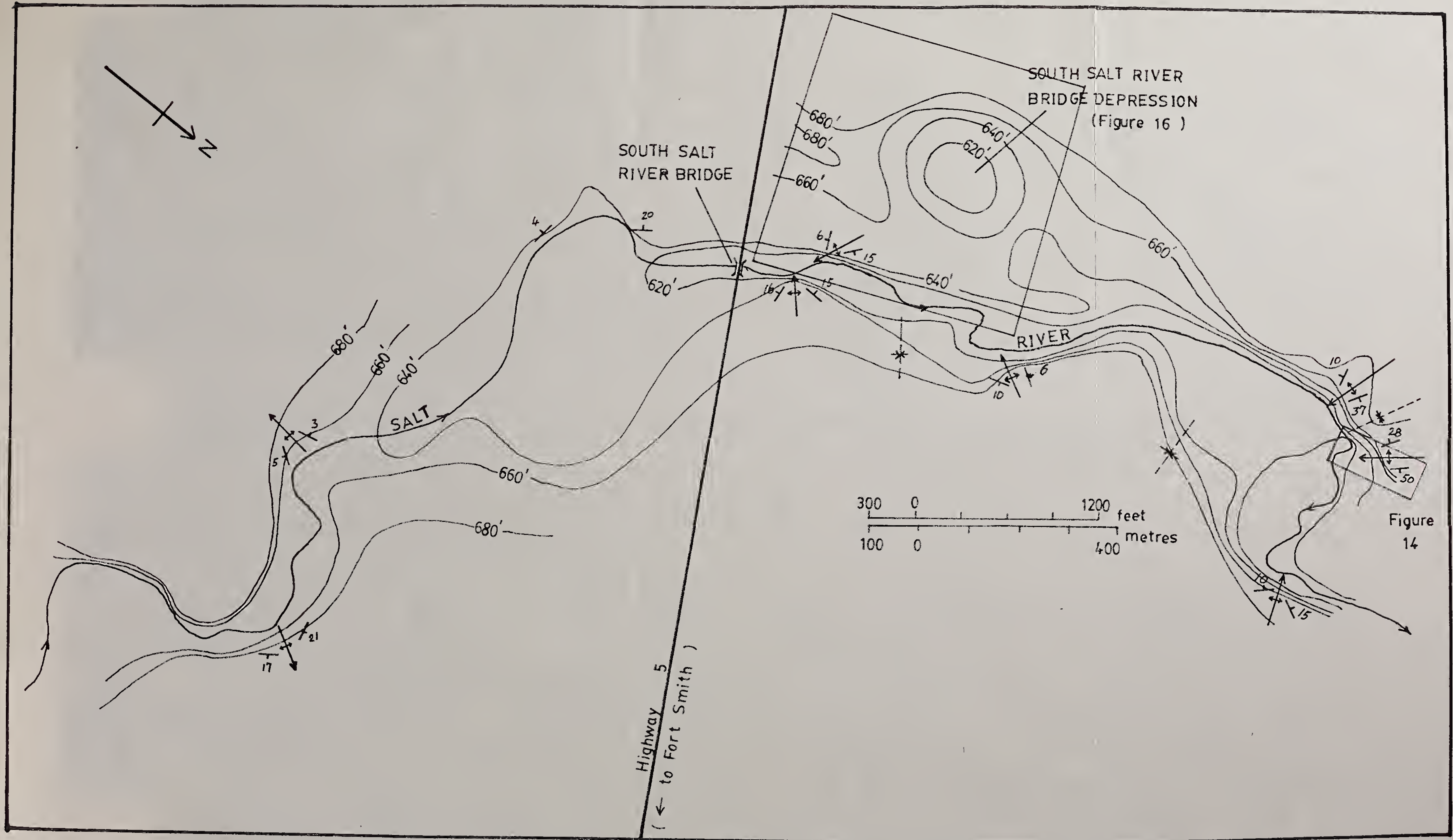


FIGURE 13. A SUMMARY PLOT OF FOLD AXES OF DEFORMED STRUCTURES EXPOSED ALONG SALT RIVER

Plate 3. Gentle folds of Member C along Salt River.

Plate 4. Anticlinal structure exposed near south Salt River Bridge.

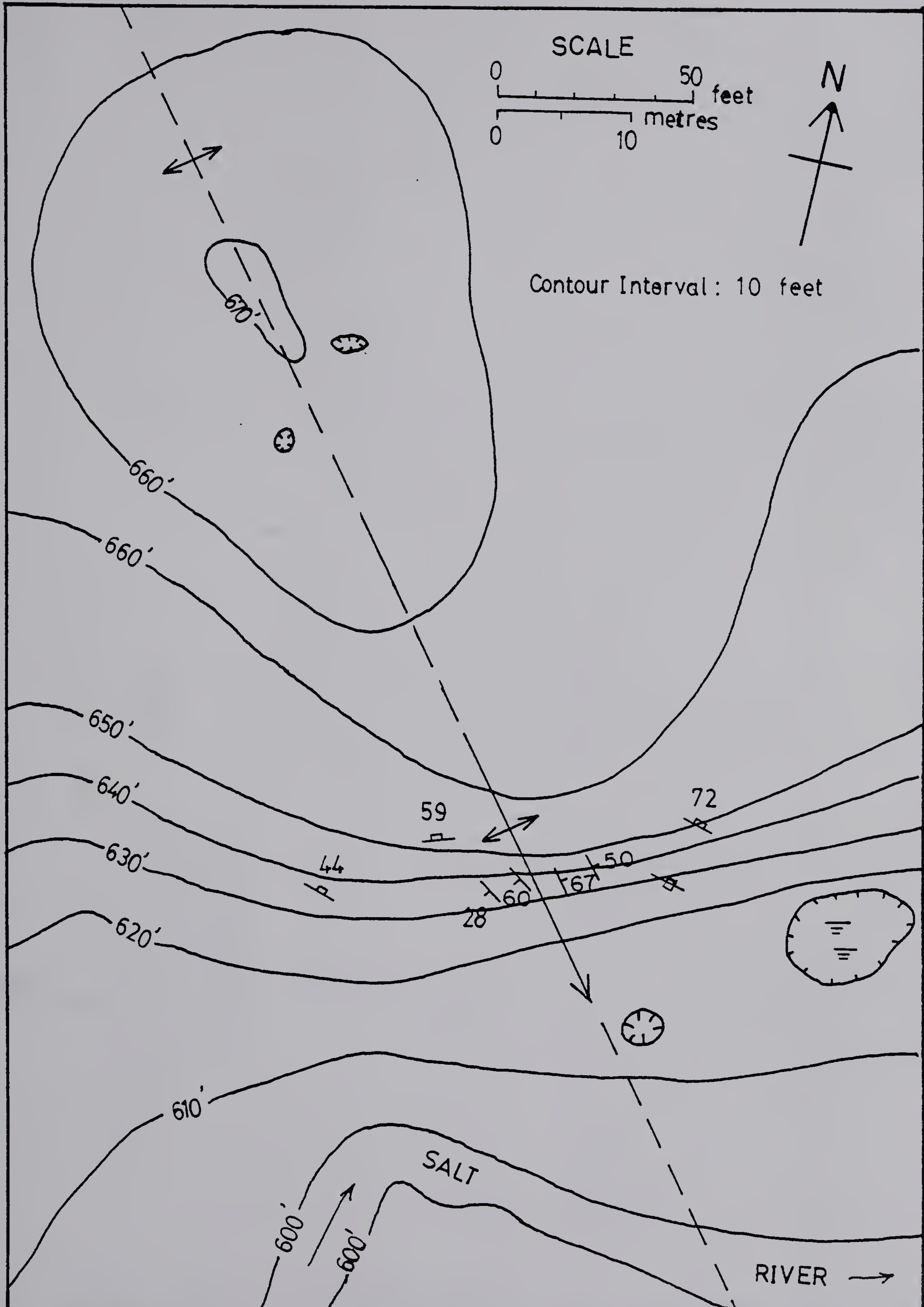
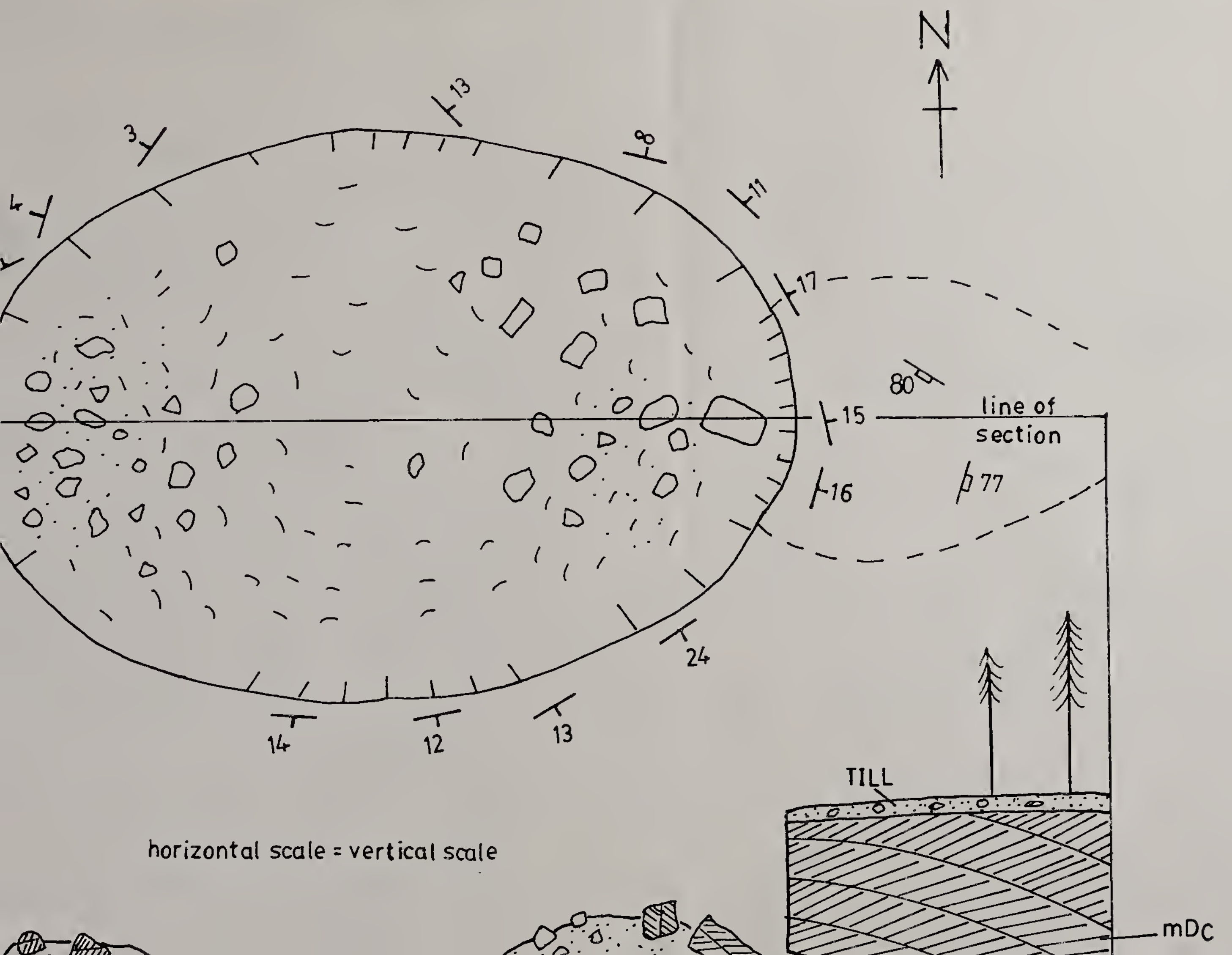


FIGURE 14. ANTICLINAL STRUCTURE ON THE WEST SIDE OF THE SALT RIVER (about 1.4 km downstream of south Salt River Bridge)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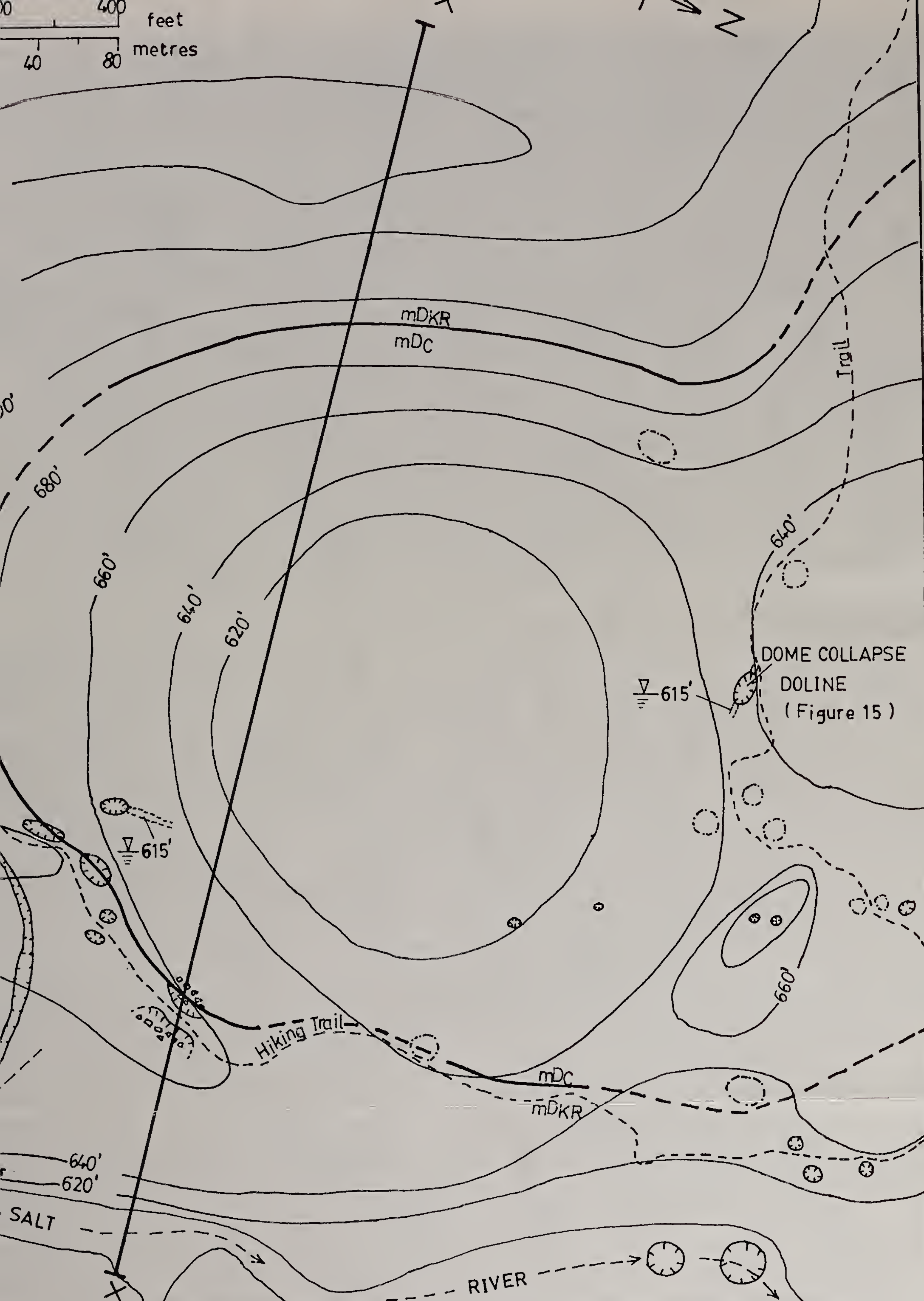
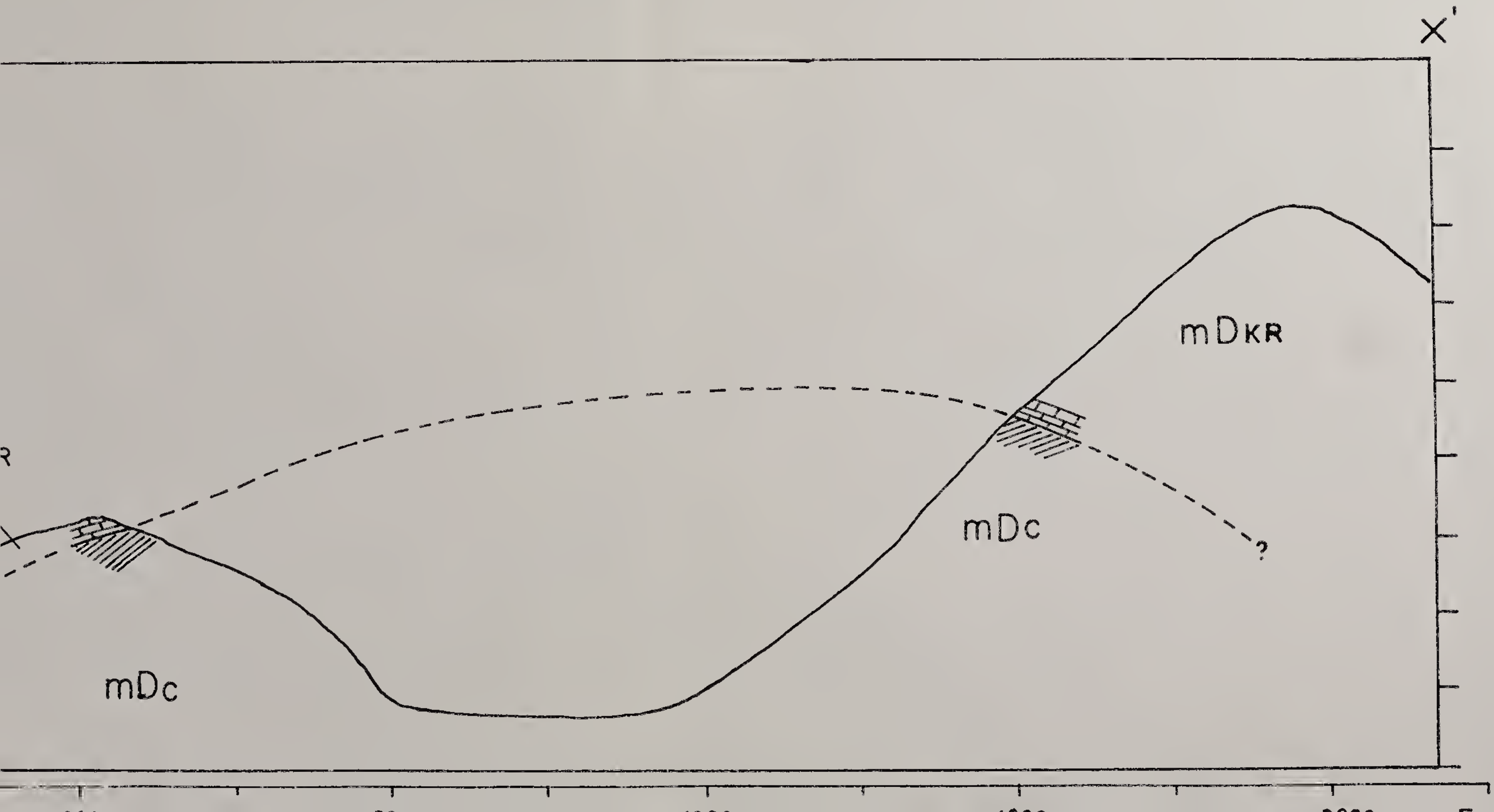


URE 16. GEOLOGY OF THE SOUTH SALT RIVER BRIDGE DEPRESSION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m) being surrounded by concentric ridges at about 1300 ft (400 m) northwest of south Salt River Bridge. The ridge is 130 ft (40 m) high measured from the base of the Depression. It is easily recognized on the stereographic aerial photographs taken in 1975 (Photo No. AS 1433-328 and AS 1433-329, scale 1 : 50,000). This is called the "south Salt River Bridge Depression" by the writer. Beds dip away from the centre of the Depression with dips decreasing from 8.4 to 1.4 degrees within a horizontal distance of 1440 ft (440 m). The boundary between the Keg River and Chinchaga Formations shown on Figures 16 and 17 is based on an altimeter survey and geologic assumption.

These features together with the arched topography and the presence of anticlinal structures along the edge of the Depression suggest that the south Salt River Bridge area had been domed. A summary plot of the fold axes and location of these deformed structures are shown in Figures 6 and 13. Figure 13 shows that there are at least 8 dome structures near south Salt River Bridge area.

Genesis

Four hypotheses are generated to explain the origin of the anticlinal structures. Each will be discussed in turn.

1. Hydration of anhydrite to gypsum

   It is known that the hydration of anhydrite to gypsum is accompanied by a volume increase of 30 % - 50 % (Deer *et al.*, 1977, p. 421; Pettijohn, 1949). Camsell

(1917), Bannatyne (1959) and Brune (1965) proposed that hydration of anhydrite had caused uplifting and formation of deformed structures in northeastern Alberta, southern Manitoba and Texas. Hydration usually takes place in anhydrite deposits after they have been uplifted close to the surface. This is because gypsum is the stable phase of calcium sulfate under low pressures and temperatures or atmospheric conditions (Bowles and Farnsworth, 1925; MacDonald, 1953; Conley and Bundy, 1959).

The presence of well-jointed overlying rocks and water bodies such as lakes and streams near the anhydritic strata are prerequisites for hydration to occur.

2. Salt intrusion

Although no beds of rock salt have been observed in outcrops and drillholes within or adjacent to the study area (Camsell, 1917; Norris, 1963, 1965; Shawinigan Stanley Ltd., 1981), facies models point out that halite crystals and beds are common in evaporitic sequences such as the Middle Devonian Elk Point evaporites of western Canada and Miocene Messinian evaporites of the Mediterranean (Kendall, 1981a, 1981b). Moreover, the brine springs with NaCl concentration over 90 % of the total solids rise along the escarpment suggest that there are rock salt deposits close to the surface (Camsell, 1917). In fact, the salt beds known as Cold

Lake Formation which is about 65 m thick, is overlain by the Chinchaga Formation and other overburden with a total thickness of about 760 m near Caribou Mountain, southwest of the study area (Shawinigan Stanley Ltd., 1980; Belyea, 1971).

Barton (1933), Nettleton (1934), Parker and McDowell (1955), and Handin and Hager (1958) performed laboratory experiments and model studies on salt movement and stated that cover of 300 m or more of sedimentary rocks can provide enough pressure to transfer the underlying salt from a solid state to a plastic state and initiate flowage (Parker and McDowell, 1955). Under this circumstance, the Cold Lake Formation may become plastic beneath the Caribou Mountains and may migrate up the southwest regional dip towards the study area because salt tends to flow from areas of high stress to areas of low stress (Romanes, 1931, p. 256). Since the overlying strata are thin near the south Salt River Bridge area, the salt may be able to overcome the weight of the rocks above, intrude upward and form domes on the ground surface.

3. Gypsum and/or anhydrite intrusions

Gypsum and anhydrite intrusions have been discovered in the Arctic Islands, northern Mexico and Persia (Heywood, 1955; Hoen, 1962; Gould and Mille, 1968; Wall *et al.*, 1961; Bush, 1929; Lees, 1931). All these movements required either a very thick overburden

(9100 m of overburden above the anhydrite beds in Axel Heiberg Island) or a compressional tectonic forces (orogenic movement during Late Cretaceous Period in the Coahuila Marginal Folded Province in northern Mexico). However, Griggs (1940, p. 1012) and Goranson (1940) stated that the creep rate of gypsum is more rapid in NaCl solution than in pure water; and Carey (1953) pointed out that the viscosity of gypsum would decrease from the original 10^{19} poises to 10^{16} poises in the presence of its own saturated solution, which is comparable or even lower than the viscosity of salt(10^{17} poises). As a result of the high NaCl concentration in the groundwater system in the study area, and the loads acting on the Chinchaga Formation caused by the overlying Upper Devonian and Cretaceous strata and glaciers (with a estimated total overburden thickness up to 670 m), a critical condition for gypsum and anhydrite intrusions may be created.

4. Ice thrusting

Deformation in drift and non-glacial sediments caused by thrust of an overriding active glacier are termed ice-thrust features (Flint, 1957, p. 88). In a map view, the ice-thrust features are linear to curvilinear, parallel and generally sharply crested ridges with the concave side facing the direction of local ice flow, and with a height of 200 ft (61 m) and an average basal width of 600 ft (183 m). In southern

Saskatchewan and eastern Alberta, exposures show that the ice-thrust ridges are composed mainly of complexly folded and faulted unconsolidated sediments of Cretaceous age, with their trends and fold axes parallel to the glacial front (Kupsch, 1962, p. 584 - 586); and in southern Manitoba, glacial drag features which occurred in the gypsiferous Amaranth Formation of Jurassic age, are composed of open and tightly closed folds with their trends approximately perpendicular to the direction of ice flowage given by bedrock striation (Wardlaw *et al.*, 1969; Bannatyne, 1971, p. 247 - 249)

The study area has been glaciated. It is possible that bedrock near south Salt River Bridge has been thrust and deformed by the ice coming from east to northeast, forming the anticlinal structures observed along the Salt River.

The first hypothesis is the most acceptable since:

a. Gypsum, anhydrite and salt domes can create anticlinal and domical structures with a diameter of at least 1.6 km. Examples: the Cambrian Salt Domes of southern Persia have an average diameter of 4.8 km (Lees, 1931), the anhydrite or/and salt domes in Canadian Arctic Islands are about 5 - 16 km across (Hoen, 1962; Gould *et al.*, 1968), the salt stocks in northwestern Germany are 2 - 8 km in diameter

(Trusheim, 1960), and the gypsum and anhydrite intrusions in northern Mexico are 14 km across (Wall *et al.*, 1961).

The anticlinal structures observed in the south Salt River Bridge area have diameters in the range of 0.006 - 0.6 km only.

b. Besides regional uplifting and downwarping, the project area has not been under any major and direct compressional tectonic stress since Devonian time (Law, 1955; Belyea, 1971; Bayrock, 1976). This is shown in the regional undeformed and near flat-lying strata within the study area.

c. Petrographic studies by Halferdahl (1960) on the gypsiferous Fort Vermilion Formation (which is lithologically and structurally similar to the Chinchaga Formation exposed in south Salt River Bridge area) at Peace Point in northeastern Alberta indicate evidences of hydration of anhydrite to gypsum.

Microscopically, all the boundaries of the anhydrite grains are marked by layers of gypsum. The anhydrite is cut by gypsum veins with sutured texture, and satin spar grows in some veins in the fibres perpendicular to the vein walls (Halferdahl, 1960). This is consistent with the textures described by Bundy (1956) which are indicative of secondary gypsum on the gypsum-anhydrite deposits in

southwestern Indiana.

d. The Chinchaga Formation is exposed or only at shallow depth (Figure 6). Gypsum would be a stable phase under low temperatures and atmospheric pressures.

e. The trends and fold axes of ice-thrust ridges should be perpendicular to the direction of local ice flow. The glacial striations exposed along Salt River show an ice flow at roughly 222 degrees (Chapter 3, Section B, p. 26); however, the fold axes of the anticlinal structures in the area point in different directions (Figure 13) and are not parallel to the local glacial front nor perpendicular to the trends of the striations. Thus, they are unlikely formed by ice thrusting.

The above arguments show that the hydration of anhydrite to gypsum should have taken place in the study area and it is responsible for the formation of the anticlinal structures.

Age

The exact age of the anticlinal structures is difficult to estimate. However, they are post-Devonian and some of them seemed to be active since the last glaciation because till is quite uniform in thickness at the crests and troughs of these structures. Moreover, no great length of time is

required to bring the hydration process to its culmination (Brune, 1965, p. 30). Redfield (1963) reported that the concrete lining of a tunnel in northern Italy has been fractured and destroyed within ten years by the surrounding anhydritic formations when they expanded in converting to gypsum.

Thus, it is reasonable to expect that these structures will continue to deform and uplift in future.

D. Faults

Description

A series of terraces are observed along Highway #5 between Rainbow Lakes and south Salt River Bridge (Plate 5). The terraces are interpreted as fault blocks which are bounded by three normal faults. These faults strike 296, 313, 329 degrees with repeated downthrow in the northeast direction forming what are known as step faults (Figure 6). The vertical displacement of these step-faulted blocks can be up to 40 ft (12 m) (Figure 18). Although these faults are difficult to follow on the ground, they can be traced on aerial photographs for about 8.1 miles (13 km), and seem to die out on the salt flat (Figure 6).

The faults are located adjacent to the anticlinal structures which appeared in the south Salt River Bridge area. Structural projection shows that one of these faults is in direct contact with the deformed structures (Figure 18).

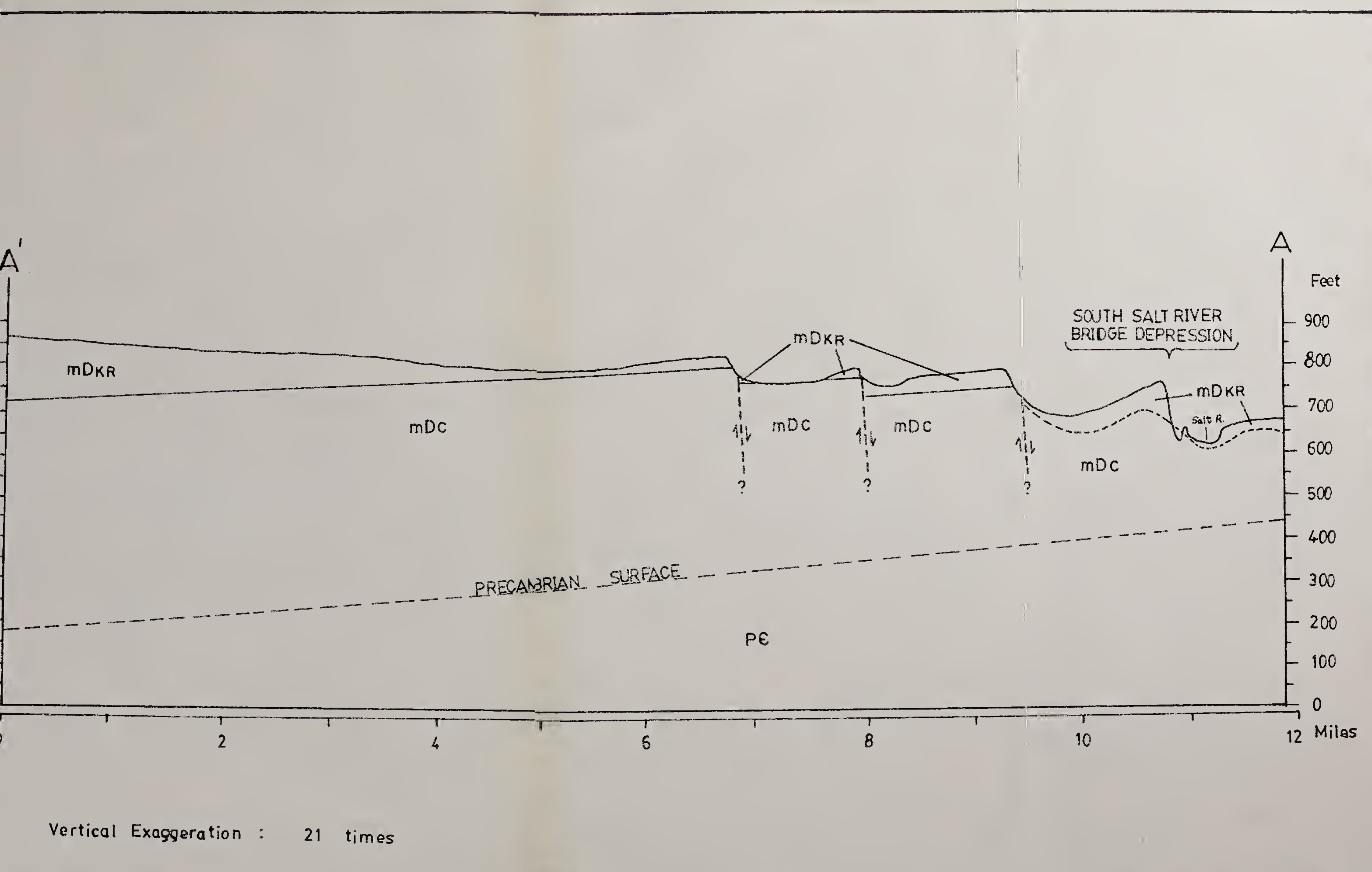


FIGURE 18. CROSS-SECTION ALONG A–A'

Plate 5. Terraces observed along Highway 5 between Rainbow Lakes and south Salt River Bridge.

Plate 6. Saggings of the limestone bed at the main falls on the Little Buffalo River.

Genesis

The faults in the study area probably are due to the combination effects of fracturing, solution and subsidence.

The presence of fresh and paleokarst features in the faulting area suggest that solution processes are and have been active near the fault zones. The intense groundwater regime in this karstic area should have caused solution, subsequent differential subsidence and faulting to take place.

Indeed, it has been reported that in Saskatchewan, the removal of salt from the Middle Devonian Elk Point Group by groundwater had produced structural depressions about 20 km long bounded by high-angle step faults (Christiansen, 1967, 1971; De Mille *et al.*, 1964). Moreover, Hackbarth and Nastasa (1979) noted that the removal of the Prairie Evaporite Formation (Elk Point Group) by solution along the Sewetakun Fault striking 338 degrees in the Athabasca Oil Sands area had caused large deformation as indicated by closed depressions with lengths 10 - 40 km and depths of 46 m in the Paleozoic rocks.

Thus, it is reasonable to postulate that there are solution generated movement in the study area which is underlain by soluble rocks.

It is certain that the step faults in the study area are post-Middle Devonian features because they cut across the Middle Devonian strata. However, there is no field evidence indicating that these faults are related or have

been influenced by the regional uplifting and downwarping since Devonian time (Webb, 1951; Belyea, 1971), although their general orientation (296 - 329 degrees) coincides with one of the two principal fracture strikes of the Precambrian basement at 315 degrees (Shawinigan Stanley Ltd., 1981).

In fact, most faults discovered outside or close to the area under discussion have a general trend of northeast-southwest. This is shown by the aeromagnetic maps which indicate a number of faults trending southwest, south of the east arm of Great Slave Lake in the Precambrian and Devonian strata which may be indicative of post Devonian movement (Douglas, 1959, p. 26). Also, the structure maps on the top of the Upper Chinchaga Formation and its equivalents, the Fort Vermilion marker, in southern District of Mackenzie and northern Alberta show the same northeast trending lineaments and had been interpreted as post-Devonian faulting (Belyea, 1971, p. 31). Furthermore, a fault which strikes 020 degrees and dips nearly vertical was found in an outcrop of Middle Devonian age at Bell Rock which is about 17 km northeast of High Escarpment (Norris, 1965, p. 87).

The normal faults in the study area do not seem to relate to any regional movement that occurred in western Canada. With the limited amount of data available, the step faults in the study area are concluded to be formed by local subsidence following solution rather than due to regional

tectonism.

Age

Faults in the area may be preglacial features based on the following reasons:

1. These faults need about 235,000 years to form by dissolving the Chinchaga Formation by groundwater (Appendix 3).
2. No post-glacial tectonic movement that can be related to the faults, appeared near south Salt River Bridge area.

The faults are possibly active at present because the vigorous groundwater system in the area would continue to dissolve the Chinchaga Formation, causing the formation of subsurface cavities and subsequent downward movement of the overlying strata into these cavities. Subsurface dissolution of the Chinchaga Formation is proved by springs which are saturated with calcium sulfate and which discharge along the base of the High Escarpment (Chapter V, Section D). Appendix 3 indicates that the downward movement of these faulted blocks is about 0.052 mm per year.

E. Evolution of the structures in south Salt River Bridge area

1. Before the beginning of the last glaciation, it is believed that local downwarping and uplifting had occurred in the south Salt River Bridge area. Dehydration of the Chinchaga Formation into anhydrite

and subsequent hydration of the rocks back to gypsum, caused a volume increase of about 30 % which deformed the overlying Keg River Formation into anticlinal structures. Figure 19 is a theoretical model for the evolution of the structures appearing near south Salt River Bridge area.

2. Immediately after the last glaciation, a vigorous groundwater system developed, transporting water deep into the Chinchaga Formation, allowing deep-seated hydration to take place. As a result, the overlying strata are continuously uplifted and deformed into anticlinal structures with gentle to close folds (Plates 3 and 4).
3. Large and long subsurface voids are created in the Chinchaga Formation at the level of the local groundwater table which is inclined towards the escarpment (Figure 27). Voids are concentrated in this level because the water table acts as vadose stream which has solution and erosional actions.
4. When the subsurface voids removed support for the overlying strata, rock failure by normal faulting occurred. The edge of blocks or the faults seem to follow the major 150 degree joint set (Figure 6, Section B in this Chapter). This is possible because the promontory is free on three sides and jointing completes the kinematic freedom of the block. Movement due to faulting increases infiltration, accelerates erosion and

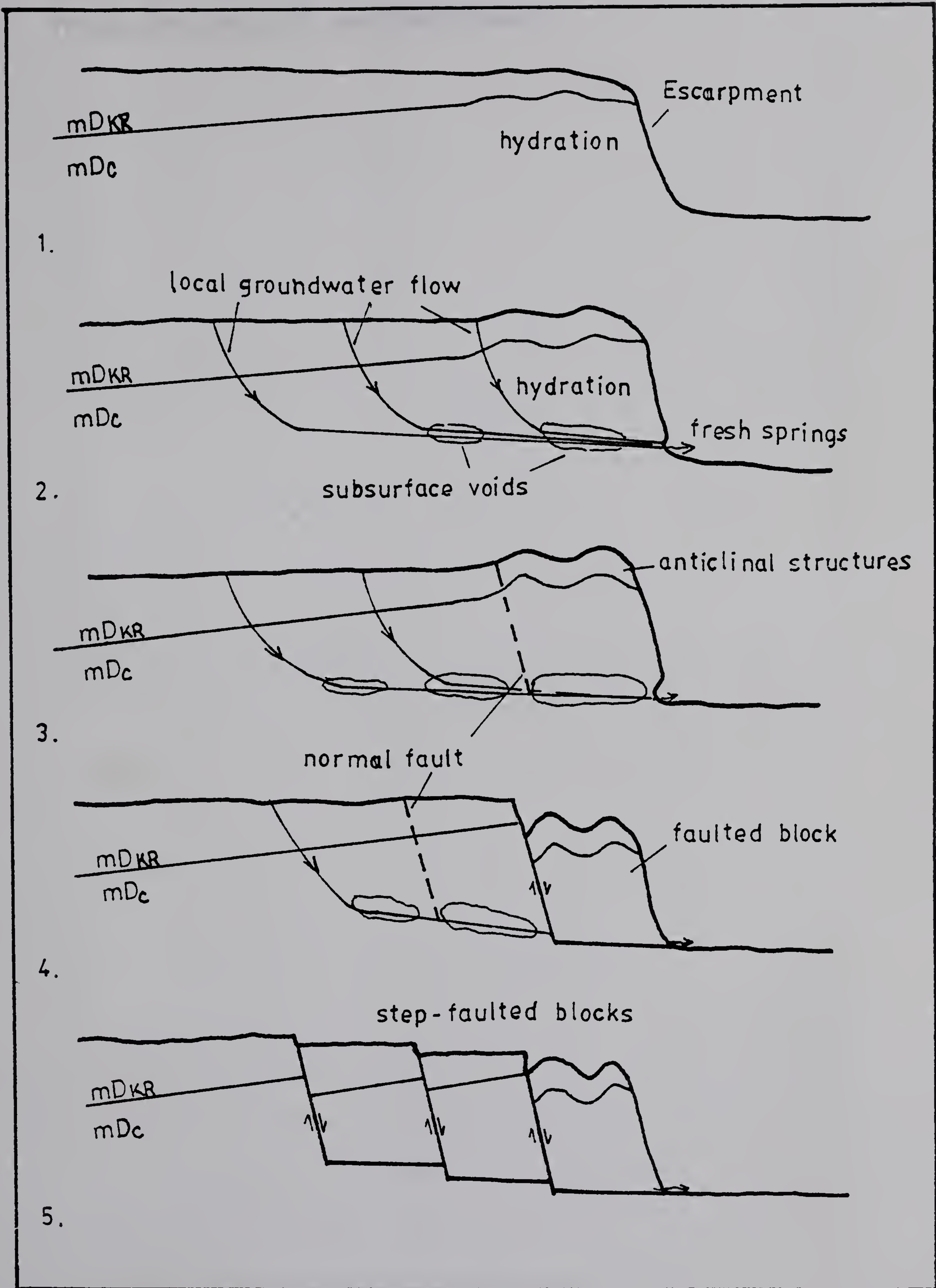


FIGURE 19. EVOLUTION OF THE STRUCTURES NEAR SOUTH SALT RIVER BRIDGE

frees the face of another block.

V. GEOMORPHOLOGY

A. Introduction

The distinct geomorphological features in the study area are:

1. Karst landforms
    a. Dolines
    b. Caves
    c. Karst valleys
2. Springs and groundwater system
3. Escarpment
4. Salt flat

An internal drainage pattern has developed in the area because of the well-jointed bedrock and low precipitation (315 mm annually) in the area. In addition to the Slave River, the Salt and Little Buffalo Rivers are the main streams responsible for fluvial erosion on the surface. Aeolian action is minor at present due to the dense vegetation developed after the glaciation.

The geomorphological features in the study area are mainly due to the extensive solution of the gypsiferous Chinchaga Formation. It is true that many large and long caves with underground streams are seen within the Chinchaga Formation; for instance, the Walk-in Cave, and other anticlinal caves within the south Salt River Bridge Depression (Figures 7 and 16). There are more collapse features than solution features in the Keg River Formation;

for example, the Fire Tower Doline and the Little Buffalo Falls (Figure 7). In fact, the study area is actually a gypsum karst area. In gypsum and anhydrite, although the solution chemistry is simpler, the landforms which result from solution erosion are generally similar to those found in limestone and dolostone (Wigley *et al.*, 1973, p. 114).

Unlike limestone karst, there are relatively few studies of gypsum karst in North America (Goodman, 1954; Olive, 1957; Ozoray, 1975; Wigley *et al.*, 1973).

B. Karst Landforms

Karst is defined as a solution-controlled landform type, characterised by an exclusive surface morphology, subsurface drainage and collapse features. It is specifically developed in calcareous rock (Sweeting, 1973, p. 333).

Due to the different lithologies and distribution of the underlying bedrock, there are two distinct geomorphological belts developed in the project area.

1. A belt of gently sloping closed depressions :

   This belt appears at and behind the escarpment where the Chinchaga Formation exposed on the surface. It is about 0.5 - 5 km wide and trends parallel to the escarpment, that is, approximately northwest. The belt is characterized by a hummocky topography with numerous solution dolines but relatively fewer collapse dolines, sloughs and karst valleys. Topographically, it resembles

hummocky disintegration moraine which has randomly orientated chaotic mounds and pits (Sugden and John, 1976, p. 254). However, the till is only 3 - 6.5 m thick (Appendix 2), the oriented depressions with their long axes usually parallel to the major joint systems of the area, and thick layers of bedrock exposed in some of these depressions exclude the possibility of the belt being a hummocky disintegration moraine.

2. A belt of steep sloping closed depressions:

   This belt occurs immediately behind the former belt where the Keg River Formation cropped out and has relatively less relief. It has a width ranging from 2.0 - 13.0 km and trends roughly northwest. The belt consists of collapse dolines, swallow holes and karst valleys (uvalas). Figure 9 shows the locations and trends of these two belts.

The karst landforms of the study area are described and interpreted as follows.

Dolines

The area has two kinds of dolines, viz., solution and collapse.

Solution Dolines

Solution doline is defined as an oval- to elliptical-shaped closed depression which is formed by dissolution and the enlargement of fissures, resulting

in lowering and settlement of the surface (Sweeting, 1973, p. 46). These solution dolines usually are 9 - 30 m in diameter and 3 - 6 m in depth with their walls inclined at about 35 degrees from the horizontal (Figure 20). Some coalesce and form compound features such as the 8-shaped dolines described by Ozoray (1975). Rainbow Lakes in the south central part of the study area probably resulted from the connection of several particularly large solution dolines (Figure 7).

Generally, the solution dolines form 25 - 35 % of the land surface where the Chinchaga Formation is exposed or only covered by a thin mantle of overburden. They are active and recent features as shown by the tilted immature spruce trees inside these solution dolines along the High Escarpment.

Collapse Dolines

Collapse dolines are the result of collapse of the roofs of large caves which lead to the formation of a near-circular closed depression with cliff-like walls and cones of debris of fallen materials (Sweeting, 1973, p. 64 - 67). Their size varies from a diameter of about 50 m and 25 m deep with overhanging walls (the Fire Tower Doline, Figure 21, and the Walk-in Cave Doline, Figure 22) to about 15 m in diameter and 5 m in depth (the Dome Collapse Doline, Figure 15).

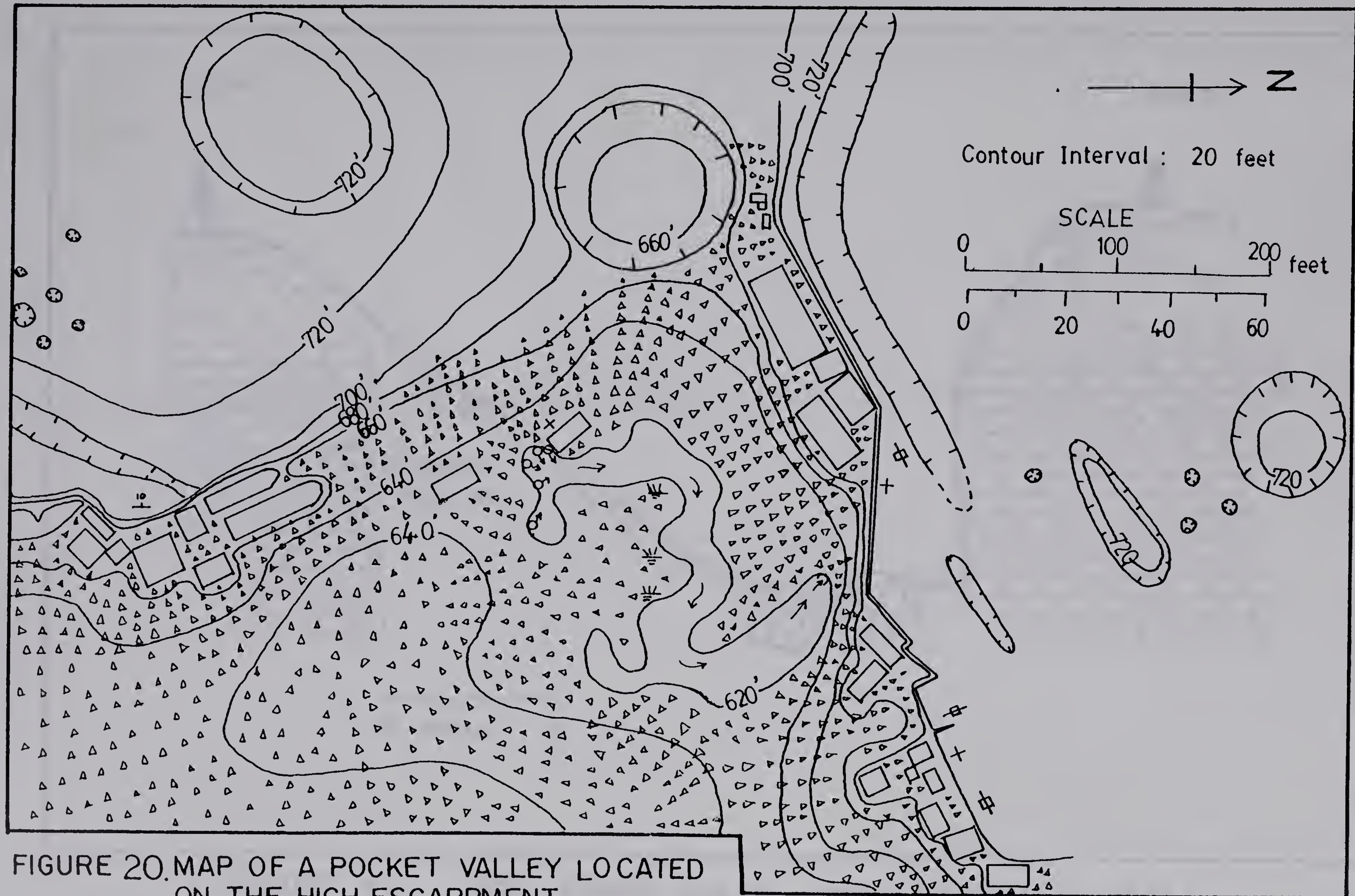


FIGURE 20. MAP OF A POCKET VALLEY LOCATED ON THE HIGH ESCARPMENT

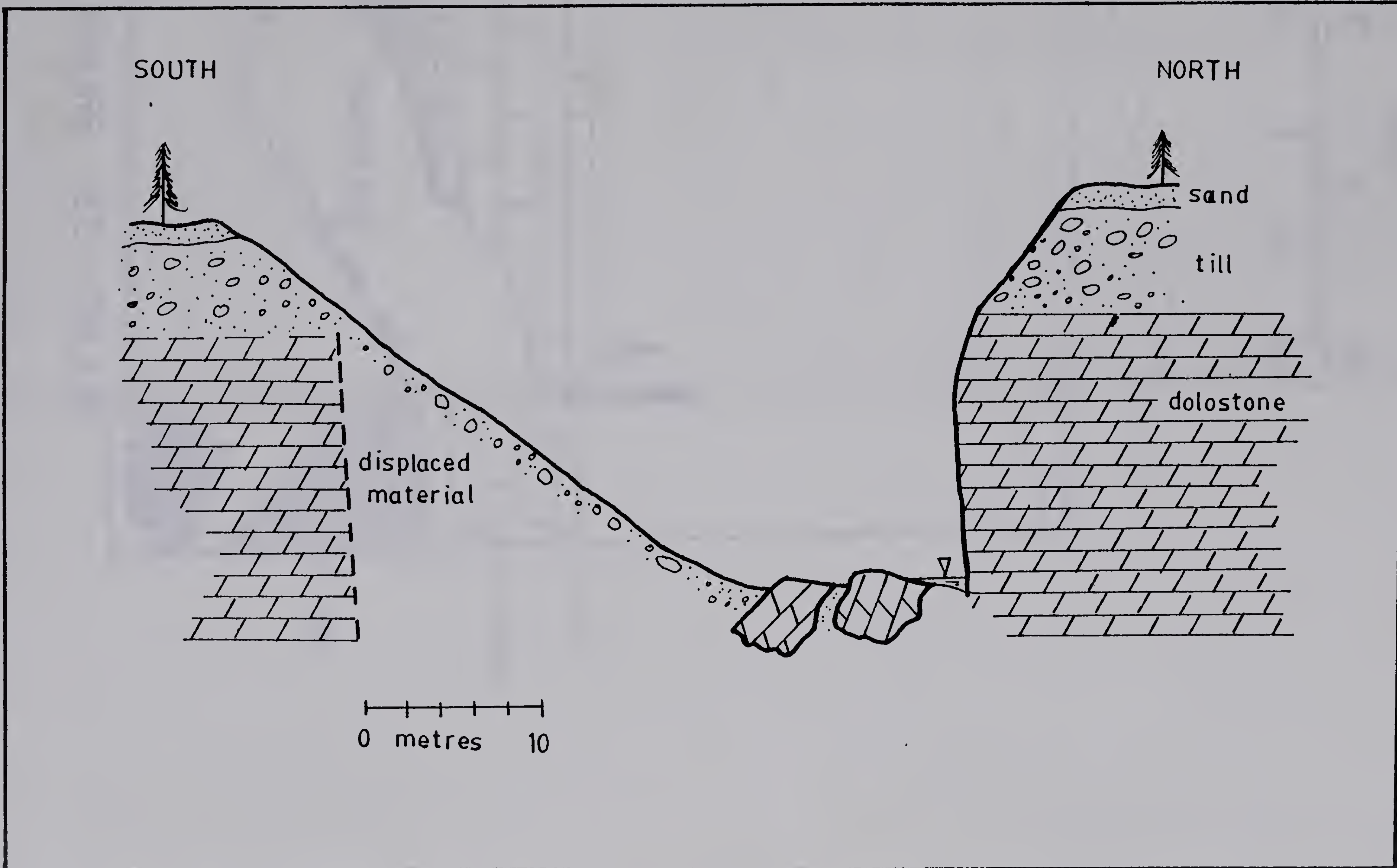
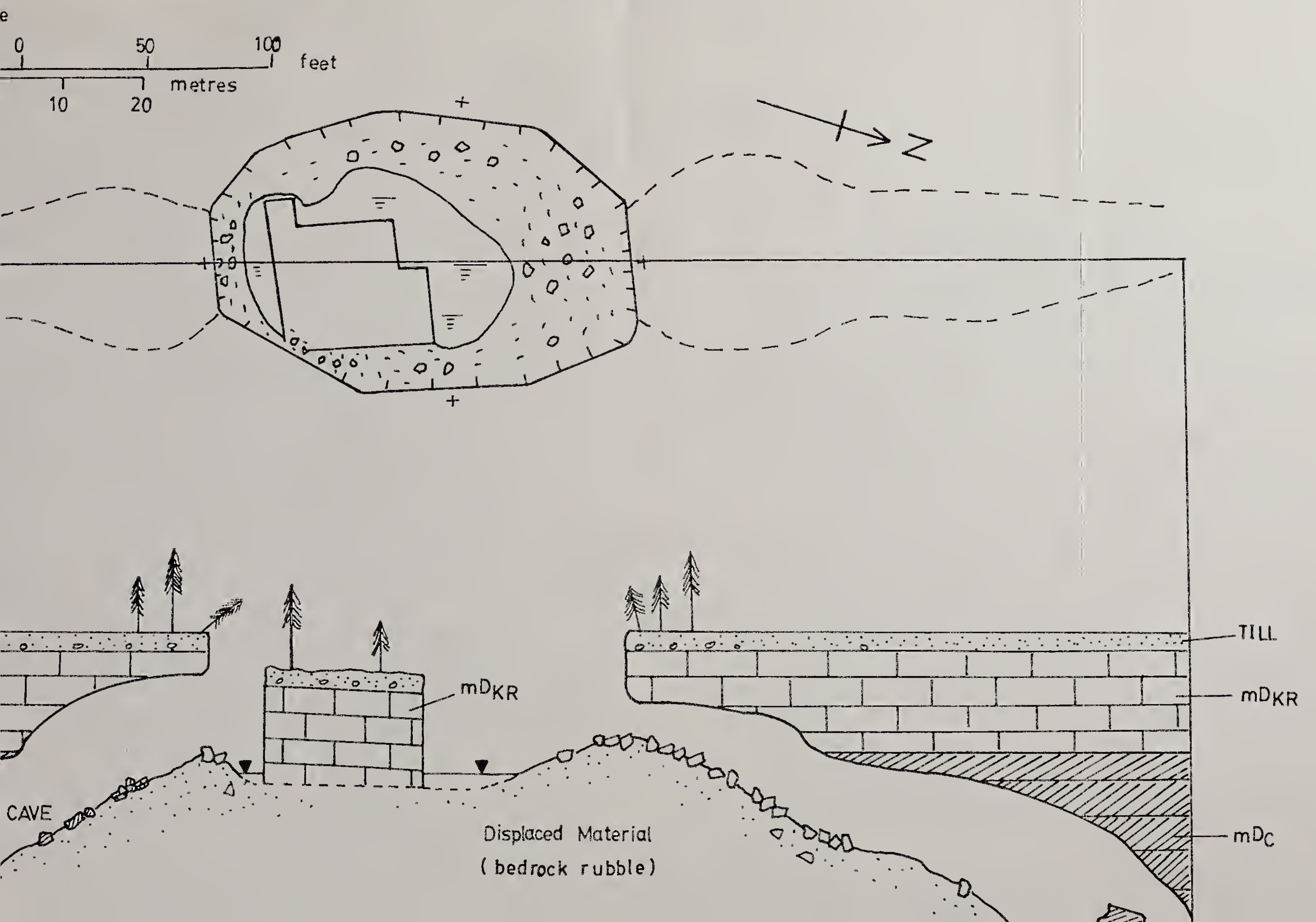


FIGURE 21. FIRE TOWER DOLINE (after Cruden et al., 1982)


e
0
50
100
feet
10
20
metres
N
TILL
mD KR
mD KR
mD C
CAVE
Displaced Material
(bedrock rubble)

On the whole, the collapse dolines covered 5 - 10 % of the area where the Keg River Formation is on the surface. However, locally they form 15 % of the land surface (Leung, 1981, p. 18). Again, they are recent features as indicated by bare overhangs, fresh and rough joints with matching asperities, and young vegetation within the discontinuities, for example, the Walk-in Cave Doline which is thought to be recent collapse structure though it appears on aerial photographs taken in 1955. Cruden *et al.* (1982) also reported that although the Fire Tower Doline appears on air photos taken in 1955 indicating it is at least 25 years old, the immature tree growth within the doline suggests the feature is not significantly older than this. In addition, Gravenor *et al.* (1978, p. 44) studied the air photographs of the area and stated that some of the solution is fairly recent as is indicated by the sand dunes which have been cut by roof collapse.

The water level observed in these collapse dolines is quite consistent and all seem to be controlled by the local groundwater system of flow. The jointed bedrock probably acts as an unconfined aquifer with its water table inclined towards the escarpment. The groundwater system of the area will be discussed in detail in Section D of this Chapter.

Figure 23 shows the stages of formation of a typical collapse doline. This is established from field

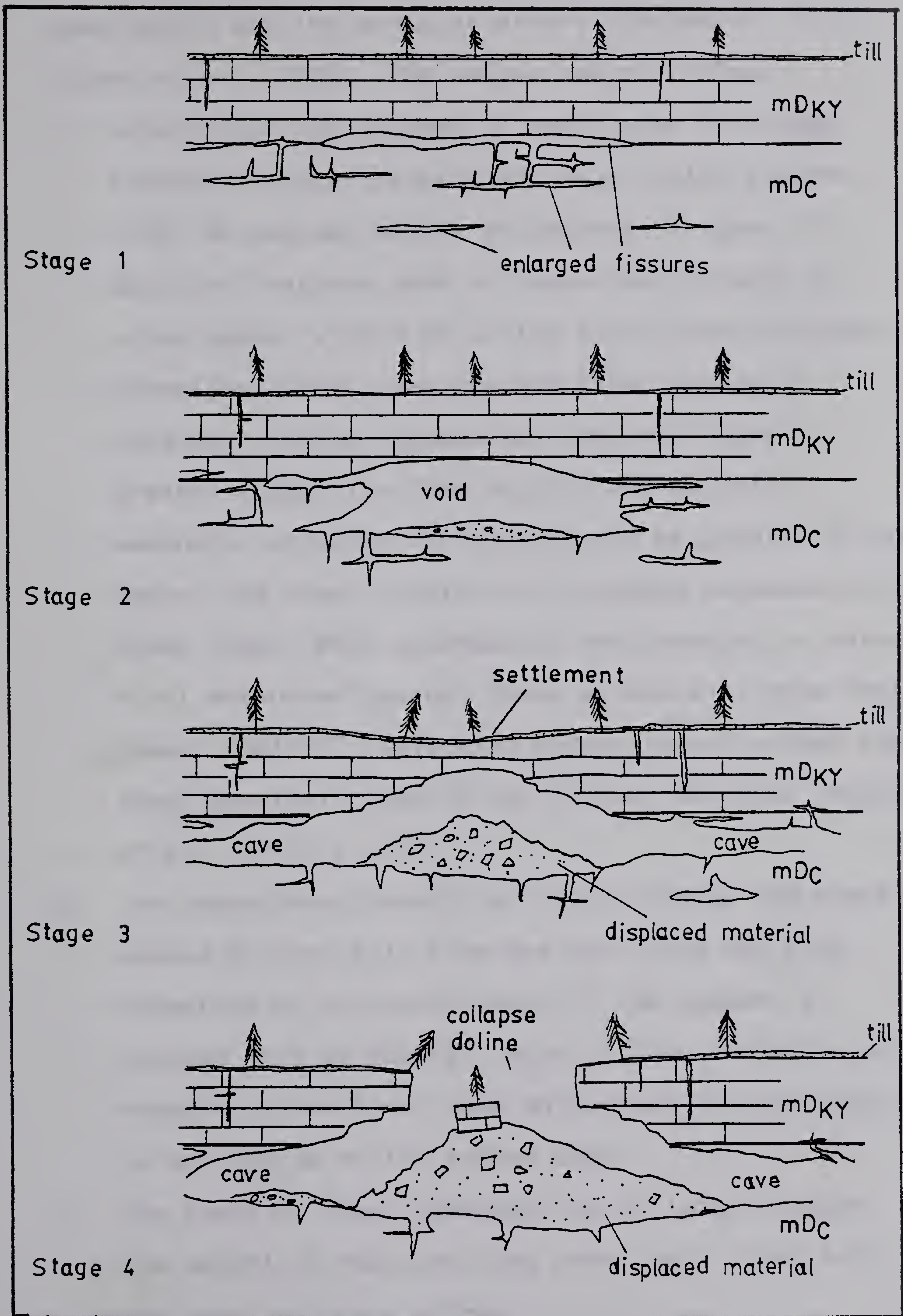


FIGURE 23. FORMATION OF A TYPICAL COLLAPSE DOLINE

observation and the works of others (Anonymous, 1979; Cruden *et al.*, 1982). The stages are as follows:

1. Solution of the exposed or underlying Chinchaga Formation concentrates along major joint systems (158, 66 degrees and 6, 86 degrees) (Figure 11). Solution features such as subsurface streams or voids appear within or at the top of the Chinchaga Formation rather than the Keg River Formation because; firstly, gypsum and anhydrite have a greater solubility than calcite and dolomite; secondly, solution activity should be greater at the upper (and lower) contact of a gypsum sequence with other rocks. When groundwater and percolation water first encounter gypsum, these waters will have their lowest "SATGYP" (saturated gypsum index) values and their greatest capacity for further solution (Wigley *et al.*, 1973, p. 127).
2. The subsurface channels or caves enlarge and migrate upward by roof fall from the overlying Keg River Formation or the overburden (if the gypsum is covered only by debris). Walk-in Cave Doline is an example of the first case while Dome Collapse Doline is an example of the second case.
3. The roofs of these openings can no longer support the weight of the overlying rock, thus, they fail and form collapse dolines.

Similar collapse structures and failure processes occurring in strata overlying soluble rock (the Prairie Evaporite Formation) in central and southern Saskatchewan have been reported by De Mille *et al.* (1964), Holter (1969), Christiansen (1967, 1971), and Gendzwill and Hajnal (1971). The Middle Devonian Prairie Evaporite (halite) grades laterally northward into the Muskey Formation (anhydrite), into carbonates of the Presqu'ile Formation in northern Alberta (Holter, 1969, p. 12), and to the lower portion of the Nyarling Formation which overlies the Little Buffalo Formation (equivalent to the Keg River Formation) in the Great Slave Lake region (Norris, 1965, p. 63; Douglas, 1970). Moreover, Wigley *et al.* (1973) described the gypsum karst in southeastern British Columbia and proposed that the formation of collapse dolines in the Middle Devonian carbonate strata (the Harrogate and Cedared Formations) by solution of the underlying Middle Devonian gypsum deposits (the Burnais Formation) had produced first cavities and later deep seated collapse of the roofs of the cavities. The Harrogate Formation can be correlated with the Methy Formation of the Clearwater River in northeastern Alberta and with the Pine Point Formation of the Great Slave Lake region (Belyea and Norford, 1967, p. 38). The Methy Formation appears to correlate approximately with the Keg River Formation in northwestern Alberta (Norris, 1963, p. 20).

Caves

Structurally, caves in the study area can be classified into two main types, viz., homocline and anticline.

Homocline Caves

Homocline caves occur in homoclinal-bedded rocks and consist of long subsurface passages, wide-span rooms, small lakes and streams. The failure of the roofs of these caves forms collapse dolines. The Walk-in Cave is a typical homocline cave in the horizontally bedded Chinchaga Formation located at the south end of the collapse doline, which is about 7.5 km behind the escarpment (Figures 7 and 22). It is about 3 - 15 m high, 10 - 14 m wide and is at least hundreds of metres long (Anonymous, 1979).

Breakdown processes are active in this cave as shown by many blocks with freshly broken surfaces and sharp edges hanging loosely from the ceiling, or piled and scattered on the floor. The presence of the underground streams in the cave also indicate fluvial erosion is active and the cave walls are being undercut continuously, leading to extensive breakdown (White *et al.*, 1969, p. 91).

Breakdown Mechanism

The breakdown sequence (Figure 23) involves:

1. A void is formed in the Chinchaga Formation due to

the considerable solution along major joint systems and also along the contact of the Keg River Formation and the underlying gypsum (also see the section concerning the formation of a collapse doline in this chapter).

2. Subsequently, the void enlarges and a subsurface passage will be developed. The passage may increase in width to a maximum of about 26 - 30 m without failure. This critical span value comes from two sources:
   a. The measurement in a large room inside the Walk-in Cave by Anonymous (1979, p. 41).
   b. The substitution of the available field data into the beam equation developed by Davies (1951) (Appendix 4).

Small differential sags obviously would occur in the bedding at the roof of the cave during its enlargement. The sags would cause bending and parting of the strata as are noticed in limestone beds cropping out on the main falls of the Little Buffalo River (Plate 6). Major breakdown probably occurred when the normal stress created in the span due to bending under its own weight exceeded the beam strength of the rock. Thus, the span fails and falls onto the floor, forming numerous broken blocks and slabs. The cycles of sagging, bending and failing repeat and cause the roof of the cave to enlarge and migrate towards the surface. Finally, a collapse

doline forms, as in the case of the Walk-in Cave.

Anticline caves

Anticline caves occur in anticlinally folded rocks and are openings that form within folded strata and have an elliptical, narrow and lens-shaped cross section.

Figure 15 shows anticline caves which appear at the east and west ends of a collapse doline in the south Salt River Bridge Depression. The cave is about 3 m high, 5 m wide and only tens of metres long. A small pond is found at the bottom of the cave which seems to connect to an underground channel. The water level taken in caves within the the Depression is at the same elevation as the Salt River; indicating they may be either connected by an underground passage or at least share the same water table. Figure 7 shows the location of the caves where the water level is taken.

Anticline caves only occur in south Salt River Bridge area.

Breakdown Mechanism

The formation of anticline caves is similar to the homocline caves. For homocline caves, the load on the span due to the weight of the beds above acts vertically on the centre of the span which creates an upward bending moment at both ends of the span. For anticline caves, the doming creates an downward bending moment at

both ends of the folded bedding.

The folding or doming of the beds produced diverging fractures from the crest of the anticline (Plate 4). As a result, the fold hinge has a higher permeability and is under a higher degree of freeze and thaw action, solution and instability. Moreover, the intersection of the fractures or discontinuities form wedges are highly unstable (for example: Dome Collapse Doline, Figure 15). Figure 24 is a stereographic plot of these joint sets and shows that a vertical line drawn through the apex of the wedge falls within the base of the wedge. According to Hoek and Brown (1980, p. 185), failure of this type of wedge can occur by fall without sliding on at least one of the joint planes. As a result, the anticline cave enlarges mainly by wedge fall.

The occurrence of the intact breccia or paleokarst features in the Middle Devonian Keg River Formation near south Salt River Bridge area (Plate 7) indicates karst processes were active at least once prior to the last glaciation.

The presence of abandoned beaches at the top of the escarpment formed by Glacial Lake McConnell indicates that the permafrost or water tables were close or at the top of the escarpment during and immediately after the

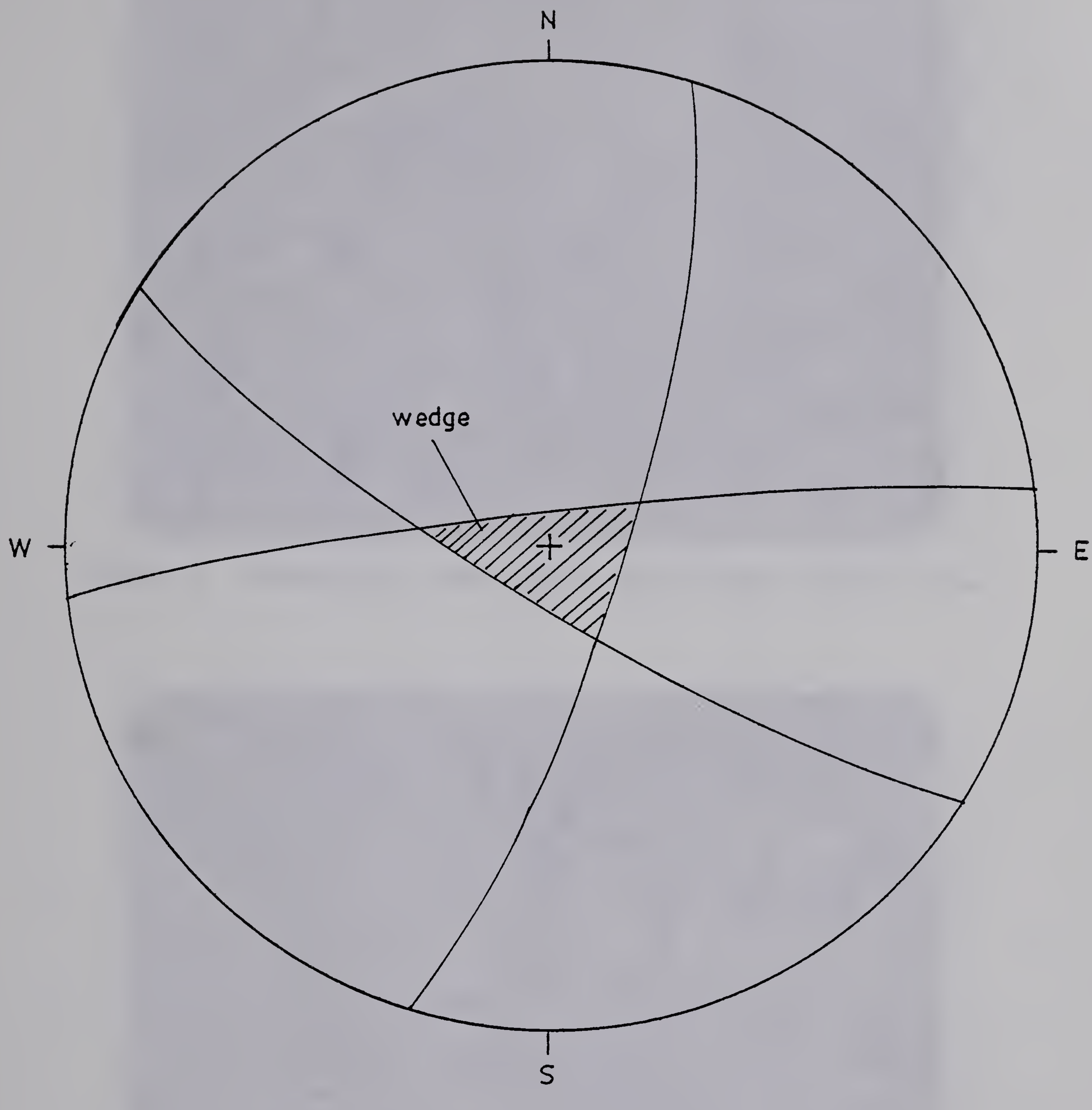


FIGURE 24. STEREOGRAPHIC PLOT OF JOINTS IN CAVE IN DOME COLLAPSE DOLINE SHOWS INSTABILITIES OF WEDGE

Plate 7. Paleokarst near south Salt River Bridge area.

Plate 8. Swallow on the Salt River.

last glaciation (Figure 9). At that time, any caves that existed would be below the permafrost or water tables. Subsequently, the retreat of the Lake caused the decay of the permafrost or/and the descent of the water table below the cave horizons. This would remove the buoyant or hydrostatic support of these caves (if any), causing major breakdown and collapse in them (White *et al.*, 1969, p. 90).

At present, the study area is in a periglacial climate and is within the discontinuous permafrost zone. The occurrence of subsurface streams and lakes, fresh fallen slabs and blocks, and the report of rocks falling while people have been in the Walk-in Cave (Anonymous, 1979) suggest that the breakdown processes are still very active. The caves are continuously attacked by seasonal flood, vadose flow and frost action.

Karst Valleys

Two kinds of karst valleys are recognized in the study area; the half-blind valleys and pocket valleys.

Half-Blind Valleys

The valleys of the Salt and Little Buffalo Rivers are flat-floored with steep sides and gentle gradients suggesting that they are meltwater channels that formed during or immediately after the last glaciation (Sugden and John, 1976, p. 307). In fact, glacial striations are

observed along the bank of the Salt River (Figure 9). Both of the rivers have become half-blind valleys locally and flow intermittently in mid-summer. Salt River originates from a lake about 22.5 km south of the Rainbow Lakes. It flows east for 30 km and then turns northwestward and flows roughly parallel to the escarpment until it nears Salt Mountain, where it turns east again and finally flows into the Slave River (Figure 2). On June 17, 1981, the Salt River with a flow of about 0.04 m/s drained totally into a swallow hole with a diameter of 5.5 m and 0.5 m deep on the river floor. The swallow hole is located approximately 200 m northwest of south Salt River Bridge (Figure 13; Plate 8). The Little Buffalo River, which had a flow of 0.3 m/s in early July entered a swallow hole 7.5 m in diameter and 4 m deep at the base of the main falls on Little Buffalo River (Figure 25). Both rivers appeared about 750 - 1000 m downstream from their points of disappearance. The rivers overflowed these swallow holes after heavy rainfalls during the field seasons.

<u>Pocket Valleys</u>

Pocket valleys are U-shaped in cross-section, flat-bottomed with steep walls and an abrupt cirque-like cliff at their head (Sweeting, 1973, p. 113). Springs commonly appear at the base of these cirque-like cliffs. Figure 20 is a sketch of a pocket valley at the High Escarpment with springs flowing at a velocity of about

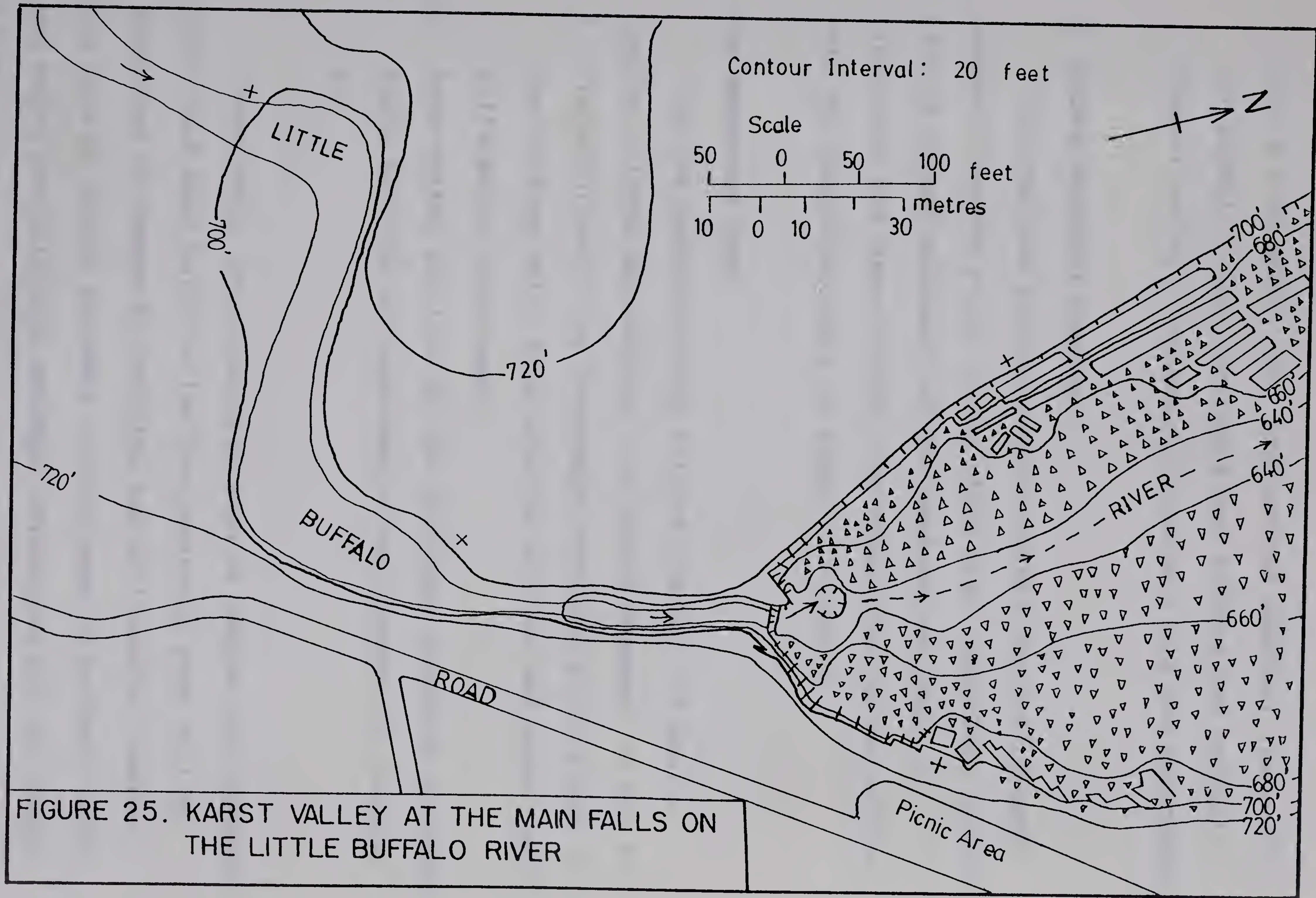


FIGURE 25. KARST VALLEY AT THE MAIN FALLS ON THE LITTLE BUFFALO RIVER

0.5 m/s in early July. The springs undercut the cliffs of pocket valleys causing slope failure and retreat. Pocket valleys are common features along the escarpment.

C. Ground Movement Hazards

The previous section clearly points out that ground movement hazards exist in the study area. According to the rate of ground movement and its destructive power to structures and human beings, the study area can be divided into two hazardous zones as shown in Figure 9.

Low Hazardous Zone

The low hazardous zone follows the belt of gently sloping, closed depressions. The ground movement is due to:

1. The solution of the Chinchaga Formation at or close to the surface which form solution dolines and cause local differential settlement.
2. Deep-seated solution of the Chinchaga Formation produces block faulting and subsidence (see Chapter IV, Section D).

Inevitably, any structures such as cabins and highways which have been built in the low hazardous zone will be subjected to damage by cracking and settlements. However, the rate of ground movement in this zone is probably slow and would possibly only endanger structures but not human lives.

High Hazardous Zone

The high hazardous zone generally follows the belt of steeply sloping closed depressions. The ground movement is due to:

1. The formation of collapse dolines in the Keg River Formation which is a sudden and catastrophic process. Its exact location is difficult to predict.
2. Deformation and/or violent "explosions" of bedrocks due to hydration of anhydrite to gypsum. Brune reported (1965, p. 29):

   > "One afternoon in 1954, 7 miles north of Moran, Texas, a farmer plowing heard an explosion and saw a cloud of dust and debris at a distance. He later found that 1,000 feet of stream channel along Deep Creek has erupted upward as much as 20 feet and scattered rock fragments over the surrounding countryside."

   Brune (1965, p. 30) believes that hydration of anhydrite to gypsum can create an uplift pressure of at least 2,000 pounds per square inch (13790 KPa).

   Hydration of anhydrite to gypsum has occurred in south Salt River Bridge area and probably the process still continues at depth at present.
3. Toppling failure along the escarpment involving rock column with volumes up to 630 cubic metres (Section E in this Chapter).

Ground collapse, hydration "explosions" and toppling failure are sudden and violent processes. They could destroy structures and endanger human lives, for example, the Walk-in Cave Collapse Doline (50 m in diameter and 25 m

deep) has the size of a campground.

Boundary of the High Hazardous Zone

The high hazardous zone is bounded easterly by the contact of the Keg River and Chinchaga Formations exposed on the surface. The approximate western boundary of this zone is based on two assumptions:

1. Collapse dolines are due to subsurface solution which are concentrated along the contact of the Keg River and Chinchaga Formations.
2. The minimum thickness of the Keg River Formation in resisting collpase at present would be exposed by collapse dolines formed recently.

The Walk-in Cave and Fire Tower Collapse Dolines are recent collapse dolines (Section B in this Chapter) and have indicated a thickness of approximately 78 ft (24 m) and 93 ft (28 m) of Keg River Formation respectively (Appendix 2, Sections 4 and 5).

In order to increase the factor of safety, the writer has arbitrary chosen 110 ft (34 m) as the maximum thickness of the Keg River Formation in which collapse dolines can be developed at present. This critical thickness is used to plot the western boundary of the high hazardous zone in Figure 9. However, this thickness will increase in future as subsurface voids enlarge and migrate upward. As a result, the western boundary will also move farther away from the escarpment in future.

The development of this area of Wood Buffalo National Park to attract more tourists in the future needs to consider safety and maintenance problems. Construction of cabins, campgrounds, highways and resort centres in the high hazardous zone (Figure 9) unless they are properly designed and located, might be subject to destruction or simply to high maintenance costs. The situation is better in the low hazardous zone; however, structures will still be damaged by slow rate of settlement and comparatively high costs of maintenance are probable.

D. Springs and Groundwater Systems

Springs

On aerial photographs, springs forming seepage zones appear along the base of the escarpment and drain onto the salt flat (Figure 9). Three springs have been examined in the field (Figure 7) and are described as follows.

1. Springs near south Salt River Bridge

   The springs rise at the base of the talus at section 6 with a flow of 0.03 m/s in early July. The water is clear but salty. The Salt River had totally drained into a swallow hole in early July which is about 1070 m upstream from Section 6. This indicates that although local relief along the Salt River channel is negligible, a local system of groundwater flow controls the recharge and discharge distance near south Salt River Bridge area.

2. Springs at the High Escarpment

The springs, which appeared at the base of screes at section 10, are about 10 m above the salt flat and flow swiftly northeast for about 70 m and disappear into a pile of blocks (Figures 20; Plate 9). They have a flow of about 0.5 m/s. Another spring found at an outcrop which is about 400 m south of section 10 has a flow of about 0.6 m/s. The water in these springs is milky and fresh, and appears to be saturated with calcium sulfate.

Figure 20 shows that there are many dolines and fissures on top of section 10 of the High Escarpment. These dolines and springs probably are local recharge and discharge areas during snow melt and heavy rainfall in spring and summer. Again, the springs in the High Escarpment are believed to be controlled by the local system of groundwater flow.

3. Springs along the base of the escarpment and on the salt flat

These springs appear at the level of playas and salt flat. They are salty and clear. Flows are difficult to measure since these springs either flow very slowly or diffuse onto the ground. For example, the springs appear at the base of the escarpment about 2.5 km southwest of the Salt River Bridge Depression (Figure 9).

Plate 9. Springs along the High Escarpment.

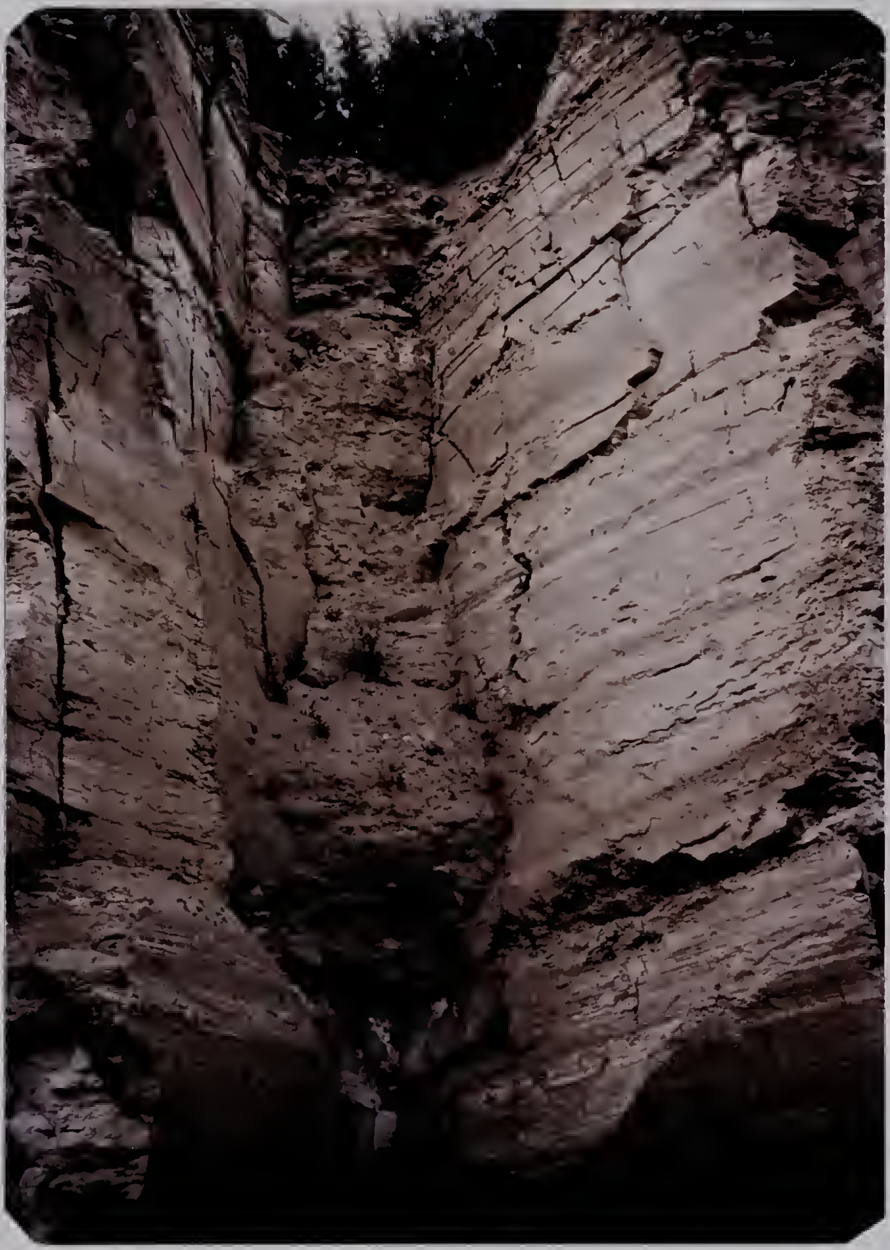


Plate 10. Wedge-shaped rock column along the High Escarpment.

Hypothetical Regional Groundwater System

At the High Escarpment, fresh springs occur at about 10 m above the salt flat where brines appear. This indicates the possibility of the existence of the discharge zones of the local, the intermediate and/or the regional systems of groundwater flow along the escarpment. Drake (1970, p. 19) has proposed that there are three hydrologic zones within Wood Buffalo National Park above the escarpment. The regional and intermediate hydrologic zones recharge near Caribou Hills and discharge on the plain between the base of the escarpment and the Slave River. The third zone recharges and discharges immediately above the escarpment (Drake, 1970, p. 19 - 23, figure 3).

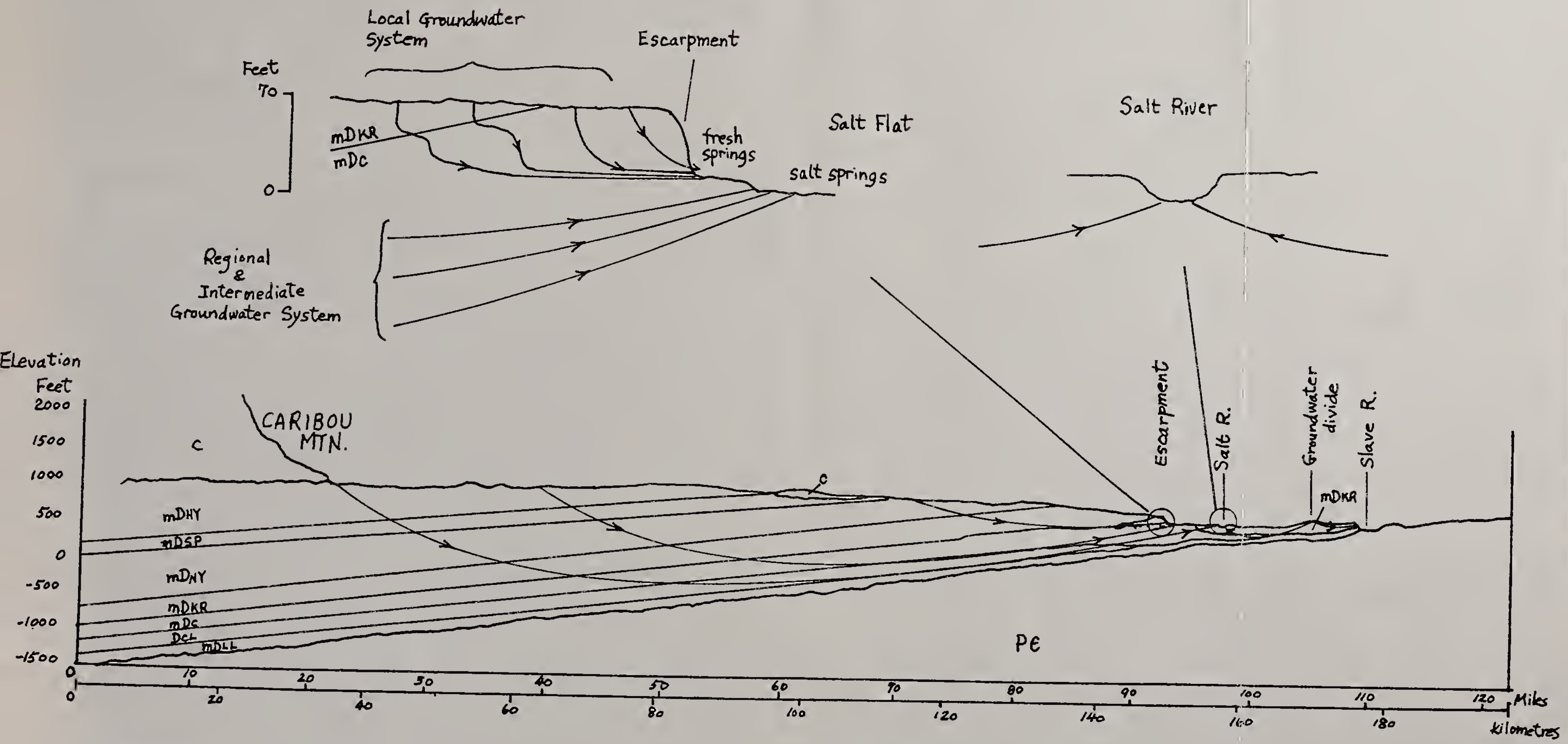
Based on Drake's hydrologic model in this area, it is reasonable to postulate that the regional and intermediate groundwater systems would flow deep enough to pass through the Cold Lake Formation which is mainly composed of salt, become brines and discharge at the base of the escarpment and the salt flat. On the other hand, the local groundwater system (or the third hydrologic zone) completes itself above the base of the escarpment and only flows through the Nyarling, Keg River and Chinchaga Formations which are mainly composed of carbonates, gypsum and anhydrite. This would cause the local groundwater system to be saturated with calcium sulfate as seen in springs at Section 10 at the High Escarpment.

Hydrologic studies (Shawinigan Stanley Ltd., 1981) indicate that the groundwater divide is about 20 km northeast of the escarpment. Thus, it is logical to postulate that the salt flat is the discharge zone for the local, intermediate and regional systems of groundwater flow (Toth and Stein, personal communication). Figure 26 shows the hypothetical relationship between the location of the fresh and brine springs, and Drake's hydrologic model within Wood Buffalo National Park.

Hypothetical Local Groundwater System

The local groundwater system of the flow is responsible for the fresh springs. This can be seen by examining the groundwater table measured in lakes, dolines and springs in the study area. Figures 7 and 27 show that hypothetical fresh water tables in dolines (recharge zone) are inclined towards the base of the escarpment where fresh springs resurge (discharge zone).

Toth (personal communication) has pointed out that the local groundwater system seems to flow from southeast to northwest and then turn to flow north to northeast when it approaches the escarpment (Figure 7). The flow direction is parallel to the orientation of the major joint system (154, 60 degrees) in the study area (see also Chapter IV, section B). In addition, Figure 27 shows that the hypothetical fresh water table has a very gentle hydraulic gradient (dh/dl) which is about 0.0016 - 0.018. And, the measured discharge



Cretaceous

C undifferentiated, Sandstone, Shale

Middle Devonian & Older(?)

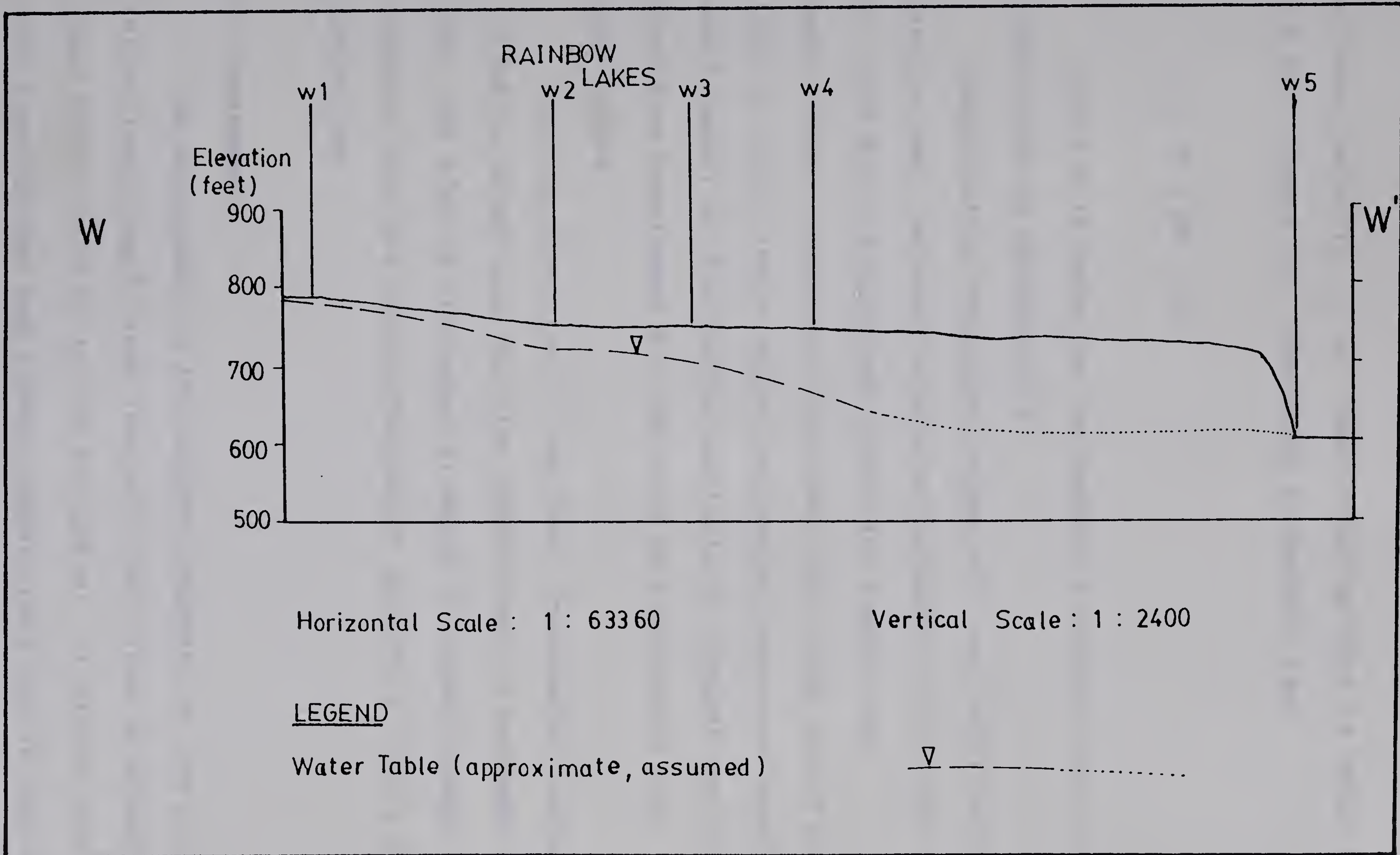


FIGURE 27 HYPOTHETICAL GROUNDWATER TABLE ALONG W - W'

or Darcy velocity (v) of these fresh springs is about 0.5 - 0.6 m/s (Figure 20). According to Darcy's Law:

$$v = -k \times dh / dl$$

where k is known as the hydraulic conductivity or coefficient of permeability .

Substituting the above values of v and dh/dl into Darcy's Law, the coefficients of permeability in the range 46 - 375 m/s are obtained, which are typical for open-jointed to heavily fractured rocks (Hoek and Bray, 1977, p. 132). The k values calculated above are based on a small amount of field data available at present and we should be considered at them only as hypothetical and qualitative.

The flow direction of the local groundwater system (SE-NW to SW-NE) and the high coefficient of permeability (46 - 375 m/s) of the material which it flows through suggests that the local groundwater system is highly joint controlled.

E. Escarpment

The escarpment is the eastern boundary of the plateau which rises steeply from the salt plain from an elevation of about 600 ft (183 m) to 735 ft (224 m). It starts near south Salt River Bridge and trends roughly parallel to the Salt River in a northwest direction towards the Little Buffalo

River Valley (Figure 9), then it gradually decreases in height northward and southward of the study area and merges with the surrounding terrain (Norris, 1963, p. 53).

Except close to the south Salt River Bridge area, the escarpment is chiefly composed of the Chinchaga Formation which is locally overlain by Members A and/or B of the Keg River Formation. For example: Section 2 is composed entirely of gypsum and is located downstream of the Little Buffalo River; and section 10 is located at the High Escarpment and exposes about 30 m of gypsum with a thin layer of carbonate beds (Appendix 2, Sections 2 and 10). The portion of the escarpment about 3.1 km west of the south Salt River Bridge is composed of relatively thick layers of the Keg River Formation overlying the Chinchaga Formation. This is probably due to faulting and folding which occurred in the area as shown in Figures 6 and 18.

French (1976, p. 147) stated that the extensive and imposing rock faces in periglacial environments might be inherited from the previous glacial period and in association with sea cliffs. Indeed, the escarpment is thought to be the former lake cliff of Glacial Lake McConnell which had flooded the regions extending from Great Bear Lake to northeastern Alberta at the end of the Pleistocene Epoch (Craig, 1965, p. 8). At present, the springs that rise along the base of the escarpment have replaced the wave action, and continue to undercut these gypsum cliffs, causing frequent slope failure and the

formation of talus.

The horizontal strata with systematic joint sets, springs undercutting, and the daily and annual freeze and thaw cycles in this periglacial environment, cause the escarpment to fail by toppling.

Toppling movements

There are two types of toppling movement occurring within the pocket valleys along the escarpment. They are the toppling of (1) Tall columns of rock topple where springs undercut the base of the gypsum cliffs, (2) Short rock columns move away from the cliff down the talus slopes. Figure 28 is a sketch of toppling failure as it occurs along the High Escarpment.

Toppling of Rock Columns

The orthogonal joint systems in the study area formed rock columns along the cliffs where the Chinchaga Formation is exposed. The columns have an average volume of 630 cubic metres, and average dimensions of 26 m (height), 7 m (length) and 3.5 m (width). Solution and undercutting along the base of the cliffs by springs and weathering have caused basal collapse and differential settlement of these rock columns. Complete undercutting is not necessary for failure as once undercutting has progressed to approximately half of the basal width, toppling then occurs (Evans, 1981, p. 81). This can be explained by the following stages (Figure 29).

1. Differential settlement due to undercutting causes the

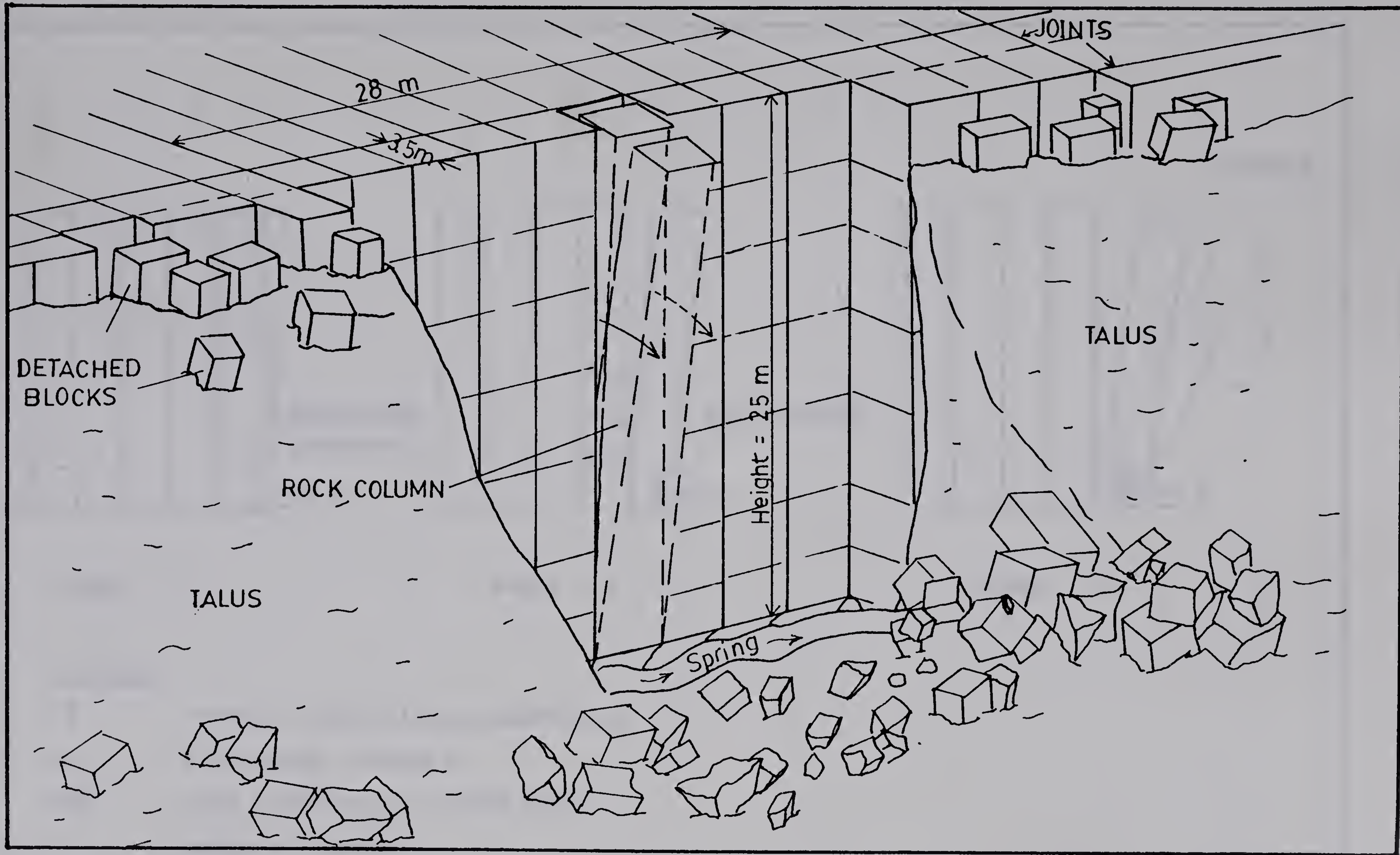


FIGURE 28. DIAGRAMMATIC SKETCH OF TOPPLING FAILURE ALONG THE HIGH ESCARPMENT

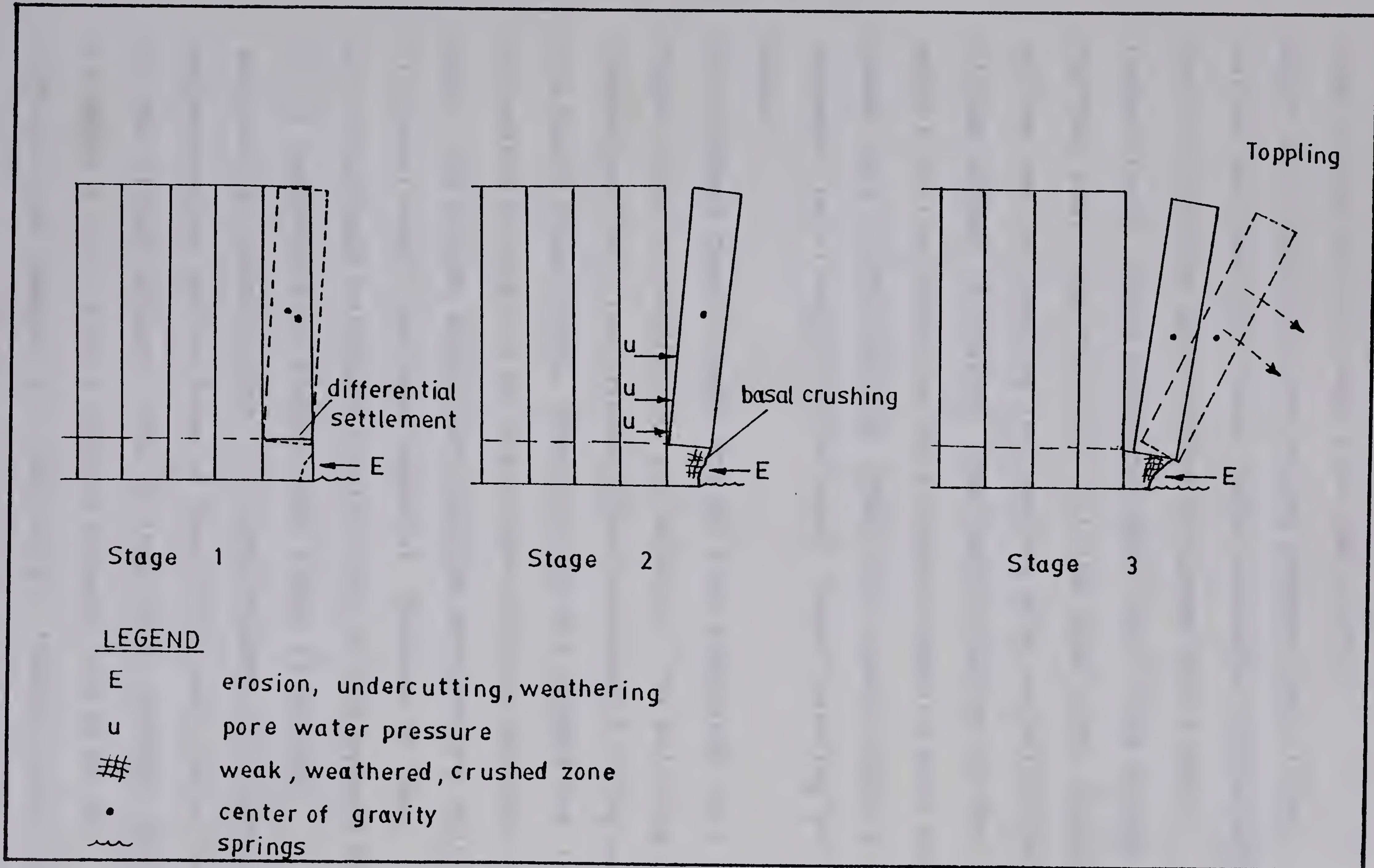
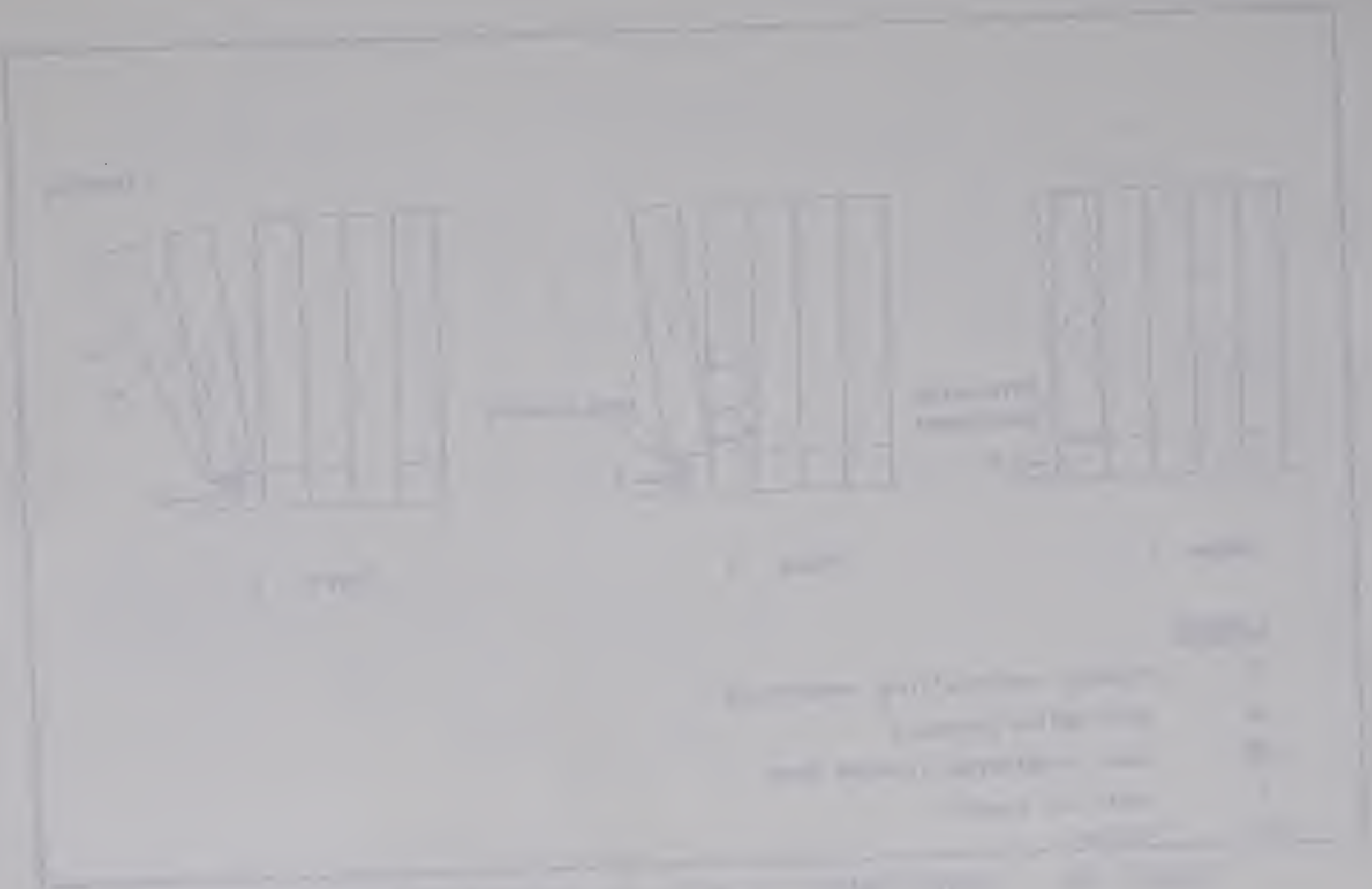


FIGURE 29. DEVELOPMENT OF TOPPLING FAILURE OF ROCK COLUMN



rock column to tilt away from the cliff.

2. Water will seep into the cracks between the tilted column and the cliff face. Water pressure (hydrostatic) and ice pressure are created in summer and winter respectively. These pressures push the tilted column farther away from the cliff. At the same time, erosion at the toe is reducing the bearing area on which the tilted column is resting. The redistribution of the weight of the column on this reduced bearing area would cause local overstressing. When local overstressing exceeds the strength of the rock, basal crushing will occur.

3. The crushed zone is weak and will be subjected to a higher rate of weathering and erosion. The settling of the column into the crushed zone increases tilting and its basal plane angle. When tilting has caused the line of centre of gravity of the column to pass outside its base, the column would have toppled even before half of its basal width has been undercut. Because of the well-developed orthogonal joint sets in the gypsum beds, it is reasonable to assume that there is neither substantial cohesion nor friction between the joint sets perpendicular to the face of the cliff and lateral faces of the tilted column. This is true as an average joint is about 0.10 - 0.60 m wide with very little or no infills (see Chapter IV, Section B). Theoretically, the basal resistance is the only force against toppling of

the rock column, and it is entirely dependent upon the normal stress due to the weight of the column. Under normal circumstances, the calculated slope of the base for an average rock column to topple is 7.6 degrees.

No column along the escarpment has been observed with tilting more than 2 degrees from the vertical. However, one needs to keep in mind the limited accuracy of the dip measurement in these nearly horizontal beds. In addition, the roughness of the faces of the rock columns and the water pressure developed between these faces and the surrounding cliffs may increase or decrease the theoretical 7.6 degrees toppling slope.

Figure 20 shows part of the cliff faces is neither parallel nor perpendicular to any of the major joint systems. The intersection of that part of the cliff with the major joint sets has formed wedge-shaped rock column (Plate 10). Evans (1981, p. 85) commented:

> "In the situation where the lateral joints are not perpendicular to the cliff face, toppling failure would be even more likely. Once even a small amount of differential settlement occurred, the horizontal stress acting parallel to the cliff face would decrease to approximately zero and no lateral shear resistance would inhibit failure."

Evans (1981) studied the toppling failures on sandstone escarpments in Illawarra Area and Gose Valley in New South Wales, Australia. He concluded that the main contributing factors to toppling in a horizontally bedded escarpment are

differential settlement by undercutting due to degradation which is aided by water pressure, weathering, the mode of joint systems, and the magnitude of the horizontal stress parallel to the cliff face. Illawarra Area and Gose Valley are under a humid sub-tropical climate while the study area has a sub-arctic continental climate. Thus, climatic condition is not a major factor for the toppling failure discussed in this thesis.

Toppling of Detached Blocks

Detached blocks have an average volume of 232 cubic metres and average dimensions of 8.1 m (height), 6.5 m (length) and 4.4 m (width). Table 1 shows the measurements of these blocks which are located in a pocket valley (Figure 20).

The floors and sides of pocket valleys along the escarpment are buried by talus which formed by the accumulation of colluvial material. This colluvium is composed of an orange-brown to brown, carbonate clayey gravel including fragments which are mainly gypsum, the remainder is limestones, dolostones, and some granitic boulders. Upon water saturation, this clayey gravel becomes a soft, sticky material with very low shear strength. The top portion of the talus has a slope of 7 - 16 degrees, and the remaining part has a slope of 17 - 47 degrees.

Detached blocks are found sitting at the top and along slopes of the talus as shown in Figure 30. These detached

Block No.	Length l (m)	Width w (m)	Height h (m)	Volume v (m^3)	Base plane angle (degrees)	Min. toppling angle $\tan^{-1}$ w/h (degrees)	Ratio w/h
1	8.53	6.71	8.23	471.1	15	39	0.81
2	6.71	6.71	8.23	370.5	15	39	0.81
3	6.40	6.71	8.23	353.4	15	39	0.81
4	15.24	6.71	8.23	841.6	31	39	0.81
5	9.14	6.71	8.23	504.7	29	39	0.81
6	6.10	5.18	6.71	334.8	14	38	0.78
7	7.32	4.57	7.01	234.5	5	33	0.65
8	7.01	4.57	7.01	224.6	8	33	0.65
9	4.27	2.13	3.66	33.3	28	30	0.58
10	6.10	6.10	10.67	397.0	10	30	0.58
11	7.32	4.27	9.14	285.7	3	25	0.47
12	4.57	1.52	3.35	23.3	9	24	0.45
13	6.40	3.96	9.14	231.6	9	23	0.42
14	3.05	3.05	7.01	62.2	13	24	0.45
15	4.27	4.27	10.67	194.5	16	22	0.40
16	8.23	3.96	11.28	367.6	11	19	0.34
17	2.74	2.44	7.62	50.9	6	18	0.33
18	4.88	2.44	10.67	127.0	9	13	0.23
19	4.27	1.52	7.01	45.5	7	12	0.21
20	2.44	1.22	7.62	22.7	16	9	0.16
21	8.84	8.23	7.01	510.0	14	50	1.20

Table 1. Dimensions, base plane angles and required toppling angles of detached blocks (at High Escarpment)

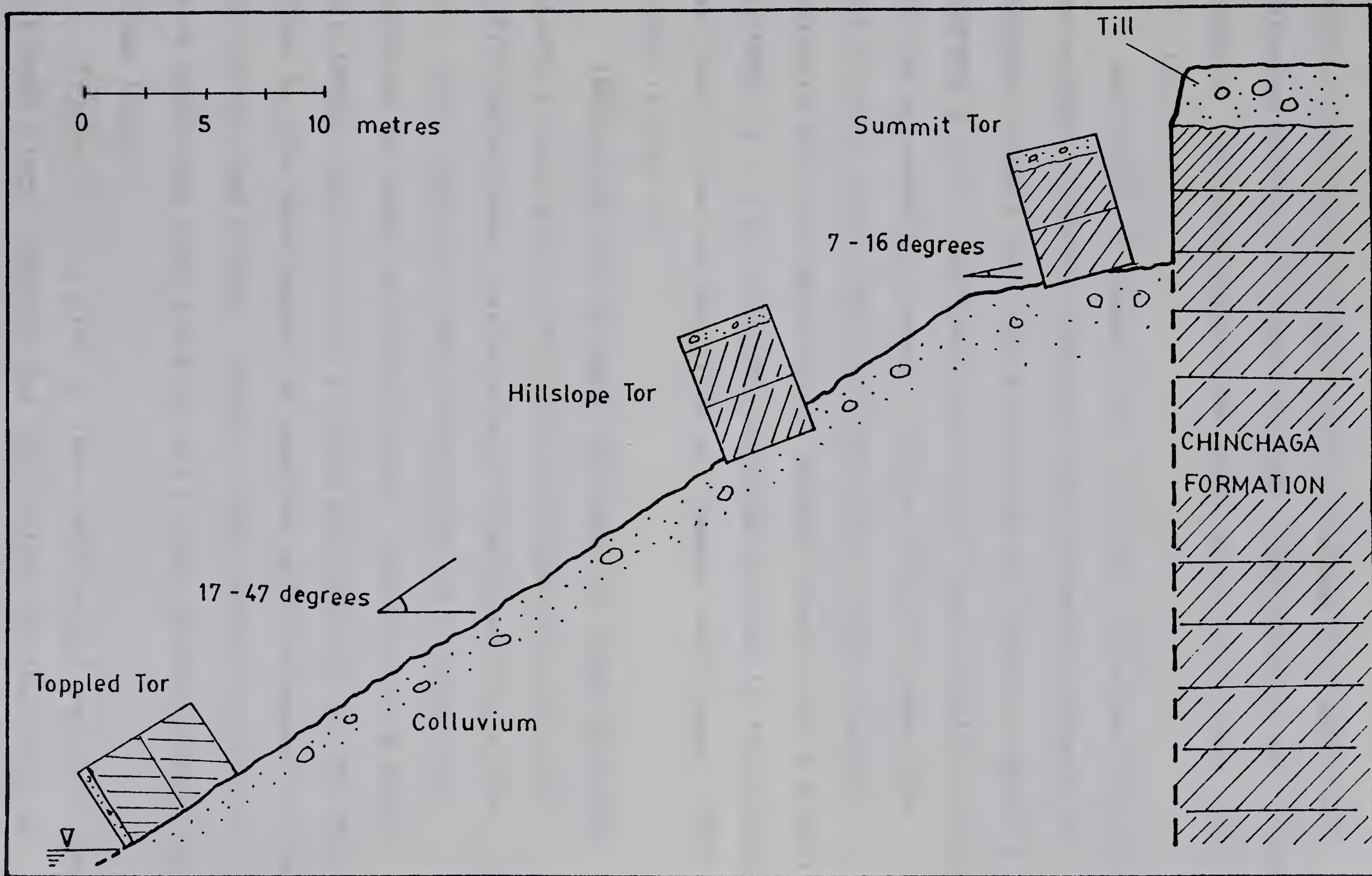


FIGURE 30. DETACHED BLOCKS (SUMMIT & HILLSLOPE TORS) ON TALUS OF THE ESCARPMENT

blocks are known as summit and hillslope tors which are detached from cliff faces through slope retreat and "frost wedging", and move downslope under gravity (French, 1976, p. 37, 134).

According to French(1976, p. 134), hillslope tors are surrounded by debris covered slopes commonly between 20 - 30 degrees; summit tors are surrounded by slopes of about 5 - 7 degrees and are located at points of high elevation relative to the surrounding terrain. Figure 20 and 25 show the position of summit and hillslope tors. These Figures indicate that the separations between these tors are quite uniform (2 - 5 m); forming features similar to the streets and rock cities as described by Simmons and Cruden (1980, p. 1305) (Plate 11).

The valley-facing basal portions of these detached blocks always settle into the colluvium and with their cliff-facing basal parts being tilted above the slope (Plate 11). This indicates that differential settlement has occurred on these detached blocks. Certainly, the basal settlement would inhibit sliding and toppling of the entire block by the development of passive earth pressure in front of the settled blocks. However, these blocks would still move downslope slowly due to soil creep process occurring on steep talus.

Figure 31 is a plot of the width/height ratio of the detached blocks against the inclination of the slope on which they are resting. The plot is based on the conditions

Plate 11. Detached blocks (tors) at the High Escarpment.

Plate 12. Salt flat with playas.

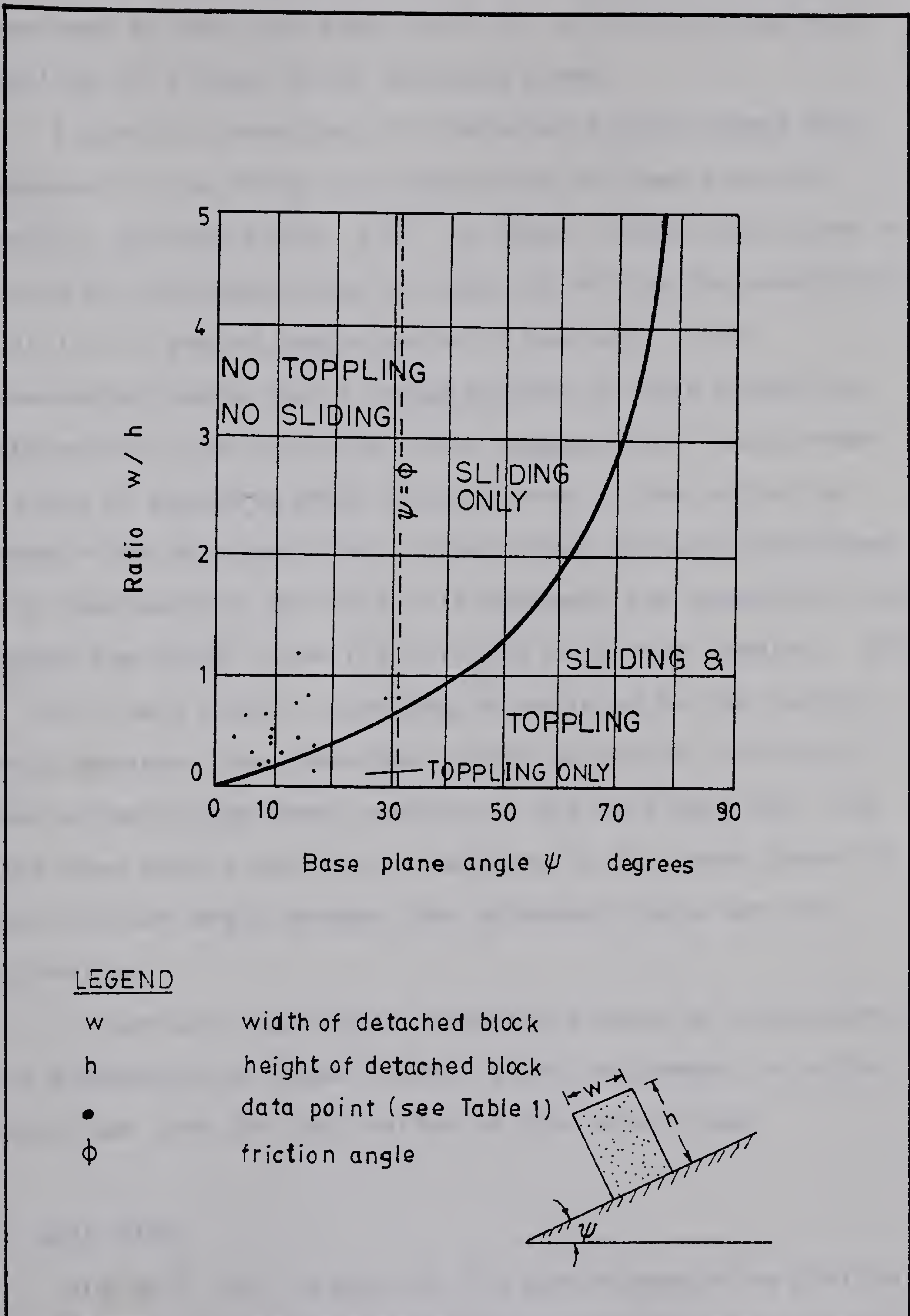


FIGURE 31. A PLOT OF WIDTH/HEIGHT RATIO OF DETACHED BLOCKS VS INCLINATION OF THESE BLOCKS

(after Hoek & Bray , 1977)

developed by Hoek and Bray (1977, p. 32) for sliding and toppling of a block on an inclined plane.

Figure 31 shows that all detached blocks except one measured in the field are stable both in theory and in reality. The one block, # 20, in Table 1 below the curve in Figure 20, indicates that it does not follow the prediction, that is, it should topple while it has not. Field observation shows that a large portion of this block has settled into the colluvium. This suggests that local-shear failure or punching shear has occurred in the colluvium beneath the detached block. Local-shear failure associated with considerable vertical soil movement and compaction, has caused the block to settle into the colluvium (Bowles, 1977, p. 115). As a result, toppling is resisted by the lateral earth pressure that develops in the colluvium in front of the valley-facing basal portion of the settled block. The plot also points out that 31 degrees is the lower bound of the friction angle between the detached blocks and the colluvium.

To sum up, toppling of detached blocks is a function of the dimensions of these blocks, their settlement into the colluvium, and the inclination of the talus slope.

F. Salt flat

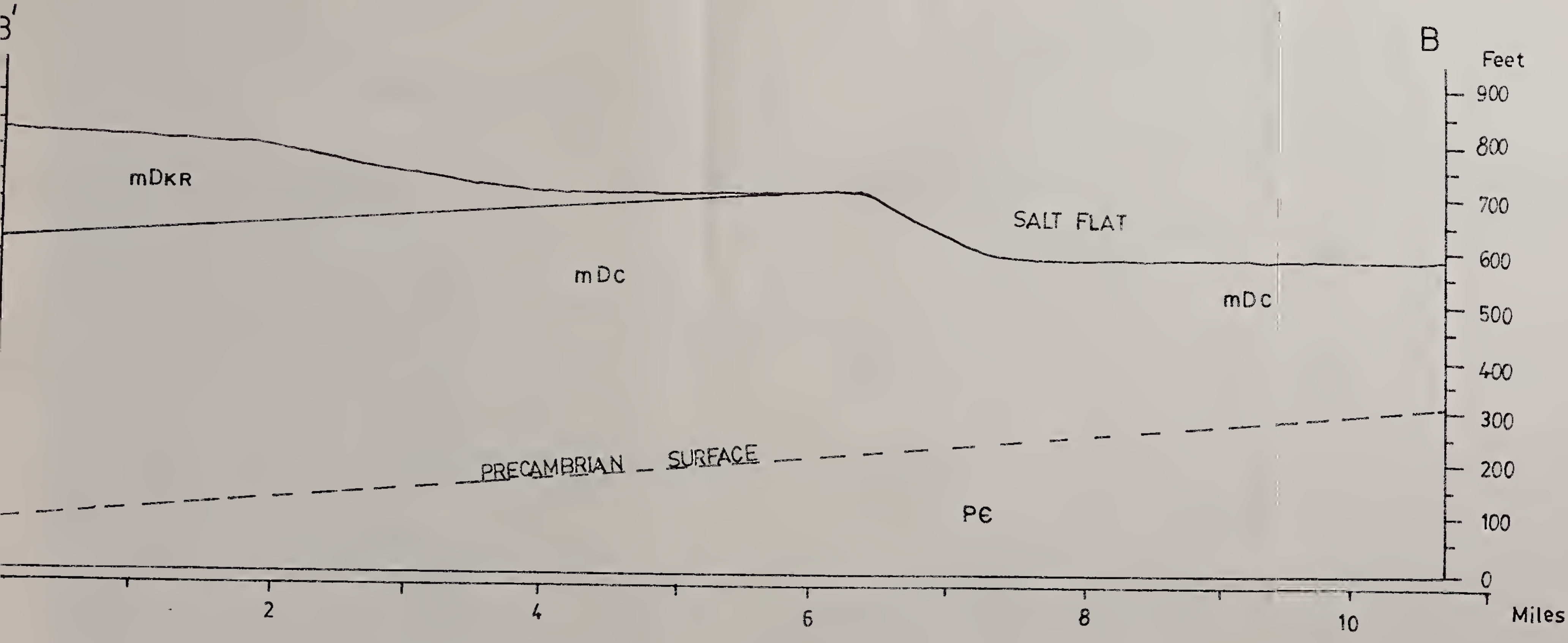
The salt flat is part of the glaciolacustrine plain at and near the base of the escarpment which has been covered by a thin layer of salt. The flat is poorly drained and

consists of many sloughs, ponds and playas which are fed by Brine Creek, Salt River, and numerous brine springs. The salt concentration in these water bodies is so high that salt continues to precipitate around playas, river channels and springs routes.

The salt is pure white and in the form of powder. It tends to form piles around boulders where the springs rise and near the edge of the playas. Plate 12 shows salt precipitates on the flat about 2.5 km southwest of the south Salt River Bridge. The salt flat can be recognized easily on aerial photos as a distinct white tone on photos. Figure 9 shows the extent of the salt flat base on aerial photograph interpretation.

In addition to the salt and glaciolacustrine deposits, gravel and boulders are scattered on the flat. The gravel is composed of fragments of gypsum, limestone, granites and gneiss. The gravel is believed to have originated from the till and the weathered bedrock at the escarpment and talus, and is being washed onto the flat by floods and streams. The boulders are mainly granites and gneiss and may be up to 0.7 m long. They are probably the ice rafted deposits of Glacial Lake McConnell.

The presence of abandoned beaches, terraces and cliffs at the edge of the salt flat indicates that it was once the boundary of a vast water body (Figures 9 and 32). The determination of the extent of Glacial Lake McConnell by Craig(1965, figure 1) shows that the salt flat was the



Vertical Exaggeration : 21 times

former lake bed of this proglacial lake. Moreover, Cameron(1922b, p. 344 - 345) described:

> "West of Fort Smith at Salt River an abandoned lake basin is distinctly shown in the salt plains lying at the foot of an escarpment facing east ... shoreline conditions are shown by narrow spruce- and poplar-clad points jutting out from the irregular face of the escarpment into the clay flats, bordered frequently by boulder pavements or shingle beaches. Low off-shore islands showing water-deposited material are also noticeable."

VI. CONCLUSION

It is believed that anticlinal structures discovered near south Salt River Bridge area are formed by the hydration and expansion of anhydrite in the Chinchaga Formation. This produced gypsum diapirs and resulted in uplifting and deformation of the beds above. Deformation fractured the rock and led to the increase in permeability and solution of the Chinchaga and Keg River Formations around south Salt River Bridge area.

Due to the local groundwater flow, large voids began to form at the level of the local groundwater table in the Chinchaga Formation near the folded area. As soon as the roofs of these openings grew too wide to support the overlying beds, they collapsed by normal faulting and produced step-faulted blocks. A model is developed to explain that the major movement occurring near an escarpment, which is composed mainly of soluble rock dipping away from the escarpment, is by block faulting (Figure 19).

The presence of intact breccia troughs (Appendix 2, Section 7) and normal faults which cut across the Middle Devonian Keg River Formation indicate that karst processes and their related movement were once active at least after the Middle Devonian Period. The sedimentation that occurred during Cretaceous time probably stopped the karst processes and they did not resume until recently, when the Middle

Devonian strata were again exposed on the surface.

Other important conclusions are:

1. The Keg River and Chinchaga Formations have at least one major joint system striking at about 60 and 154 degrees, which is approximately normal and parallel to the trend of Cordilleran orogenic belt. This suggests that the major joint system in the study area might be related to the past tectonic events such as the Laramide orogeny.
2. The presence of abandoned beaches, terraces and cliffs along the escarpment shows that the escarpment is probably the shoreline of Glacial Lake McConnell that appeared at the end of the last glaciation. At present, the springs that discharge at the base of the escarpment have replaced the wave action and continue to undercut these cliffs.

   The combination of basal erosion, frost action and the orthogonal joint systems in this horizontal-bedded escarpment have produced rock columns and detached blocks (summit and hillslope tors) that fail by toppling. The theoretical toppling angle for these rock column is 7.6 degrees. The lower bound of the friction angle between the detached blocks and the colluvium is found to be equal to 31 degrees.
3. A thin layer of till was deposited over the study area during the last glaciation. The retreat of Glacial Lake McConnell has left a vast glaciolacustrine plain between the escarpment and Slave River. The abandoned beaches of

the lake are the source for the sand dunes in the study area.

4. Geomorphologically, the study area is a gypsum karst area and can be divided into two zones or belts:
   a. Low hazardous zone or belt which consists of many solution dolines and where the gypsiferous Chinchaga Formation is exposed on the surface. The hazards that occur in this zone would damage structures and increase maintenance costs.
   b. High hazardous zone or belt which consists of collapse dolines and where the carbonate Keg River Formation is exposed on the surface. The hazards that occur in this zone may cause death and complete destruction of structures.

   The karst processes are active at present as shown by many dolines and caves with distinct and fresh collapse features.

5. The fresh and salt springs that appear at the base of the escarpment are controlled by the local, intermediate and regional groundwater systems of flow. The discharge zones of these systems are concentrated at the base of the escarpment and at the salt flat, causing the fresh springs to discharge at about 10 m higher than the salt springs.

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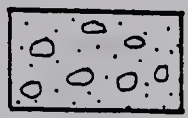
Appendix 1 : Conversion factors, lithologic column and map symbols

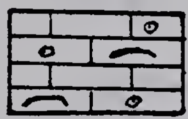
Conversion factors

1 inch	=	2.54 cm
1 m	=	3.281 ft
1 m^2	=	10.764 ft^2
1 m^3	=	35.315 ft^3
1 km	=	0.6214 mile
1 ft^3/sec	=	0.02832 m^3/sec
	=	28.32 litre/sec
1 Pa	=	1 Newton/m^2
1 MPa	=	1.0×10^6 Newton/m^2
1 KPa	=	0.145 lb/in^2
1 bar	=	14.50 lb/in^2
1 kg	=	2.2046 lbs.

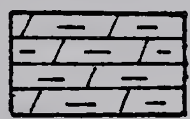
LITHOLOGIC SYMBOLS FOR COLUMNAR SECTIONS

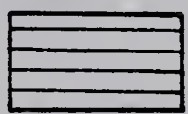

Conglomerate

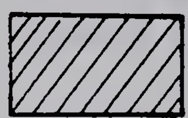

Limestone

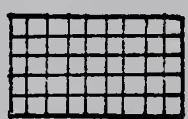

Fossiliferous limestone

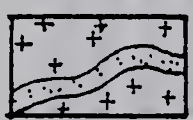

Dolostone


Argillaceous dolostone

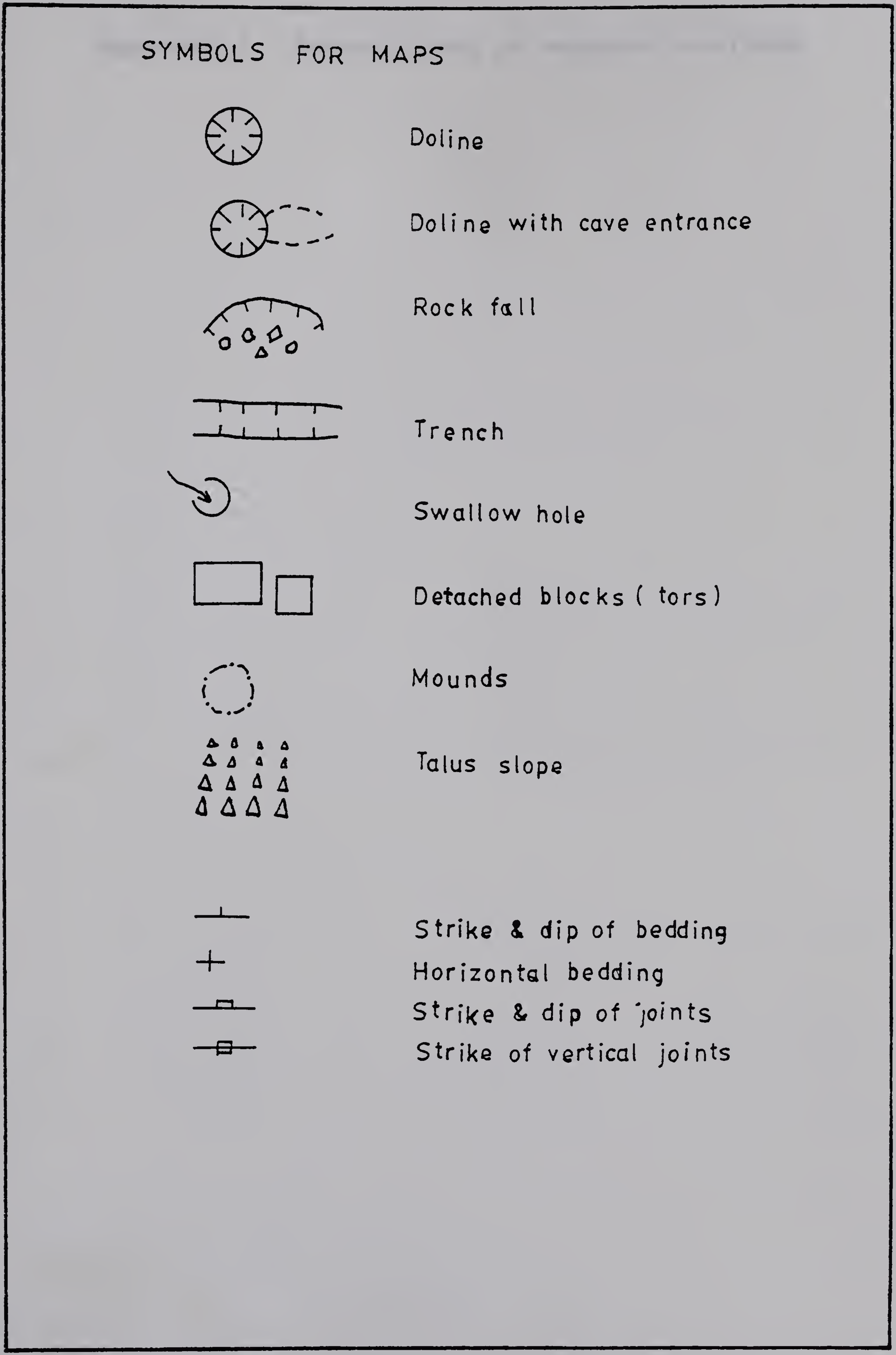

Shale


Gypsum , anhydrite


Salt

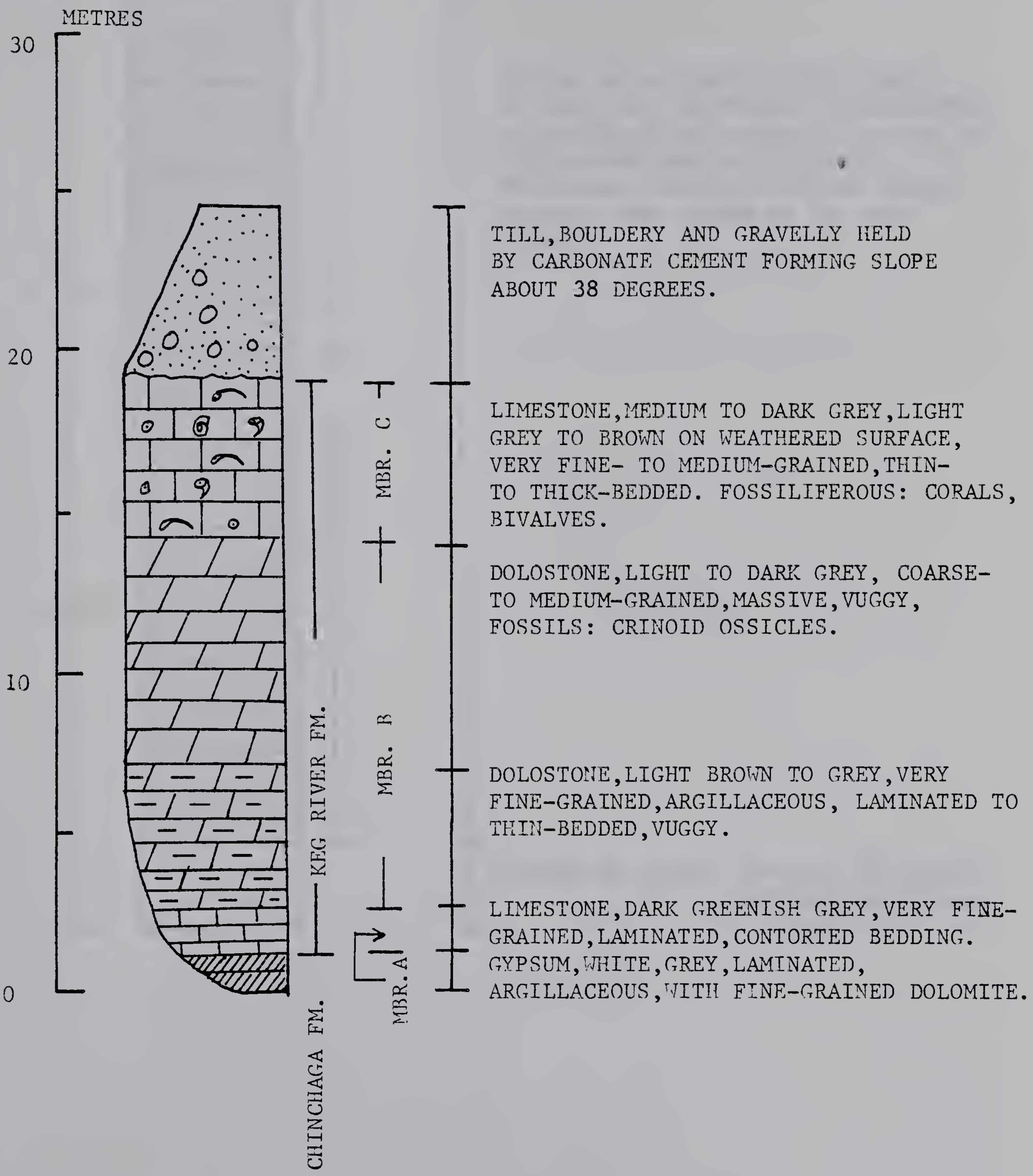

Granitic gneiss

A	Member A
B	Member B
C	Member C


SYMBOLS FOR MAPS
Doline
Doline with cave entrance
Rock fall
Trench
Swallow hole
Detached blocks (tors)
Mounds
Talus slope
Strike & dip of bedding
Horizontal bedding
Strike & dip of joints
Strike of vertical joints

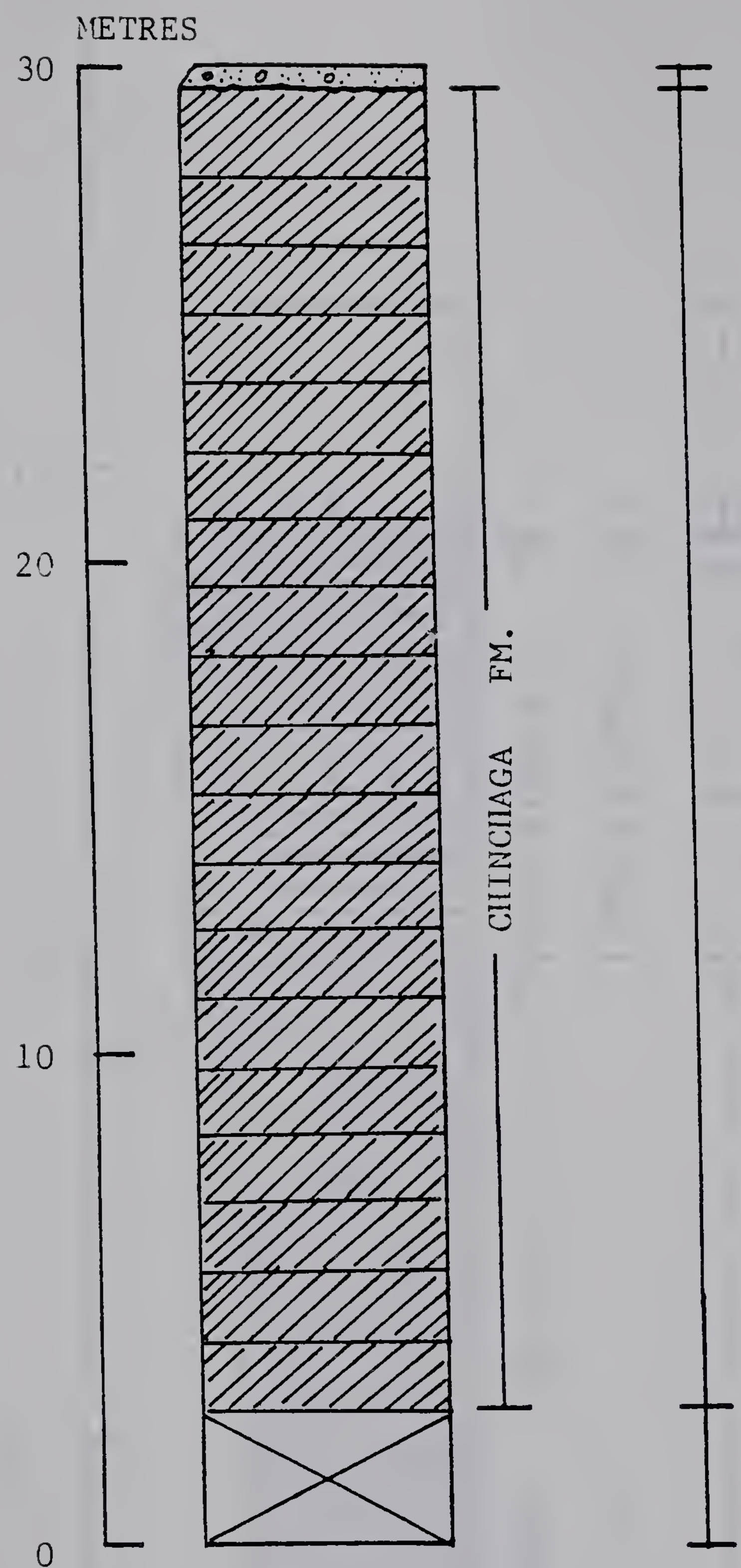
Appendix 2 : Descriptions of measured sections

SECTION I



METRES
30
20
10
0
TILL, BOULDERY AND GRAVELLY HELD BY CARBONATE CEMENT FORMING SLOPE ABOUT 38 DEGREES.
LIMESTONE, MEDIUM TO DARK GREY, LIGHT GREY TO BROWN ON WEATHERED SURFACE, VERY FINE- TO MEDIUM-GRAINED, THIN- TO THICK-BEDDED. FOSSILIFEROUS: CORALS, BIVALVES.
DOLOSTONE, LIGHT TO DARK GREY, COARSE- TO MEDIUM-GRAINED, MASSIVE, VUGGY, FOSSILS: CRINOID OSSICLES.
DOLOSTONE, LIGHT BROWN TO GREY, VERY FINE-GRAINED, ARGILLACEOUS, LAMINATED TO THIN-BEDDED, VUGGY.
LIMESTONE, DARK GREENISH GREY, VERY FINE-GRAINED, LAMINATED, CONTORTED BEDDING.
GYPSUM, WHITE, GREY, LAMINATED, ARGILLACEOUS, WITH FINE-GRAINED DOLOMITE.
KEG RIVER FM.
CHINCHAGA FM.
MBR. C
MBR. B
MBR. A

SECTION 2

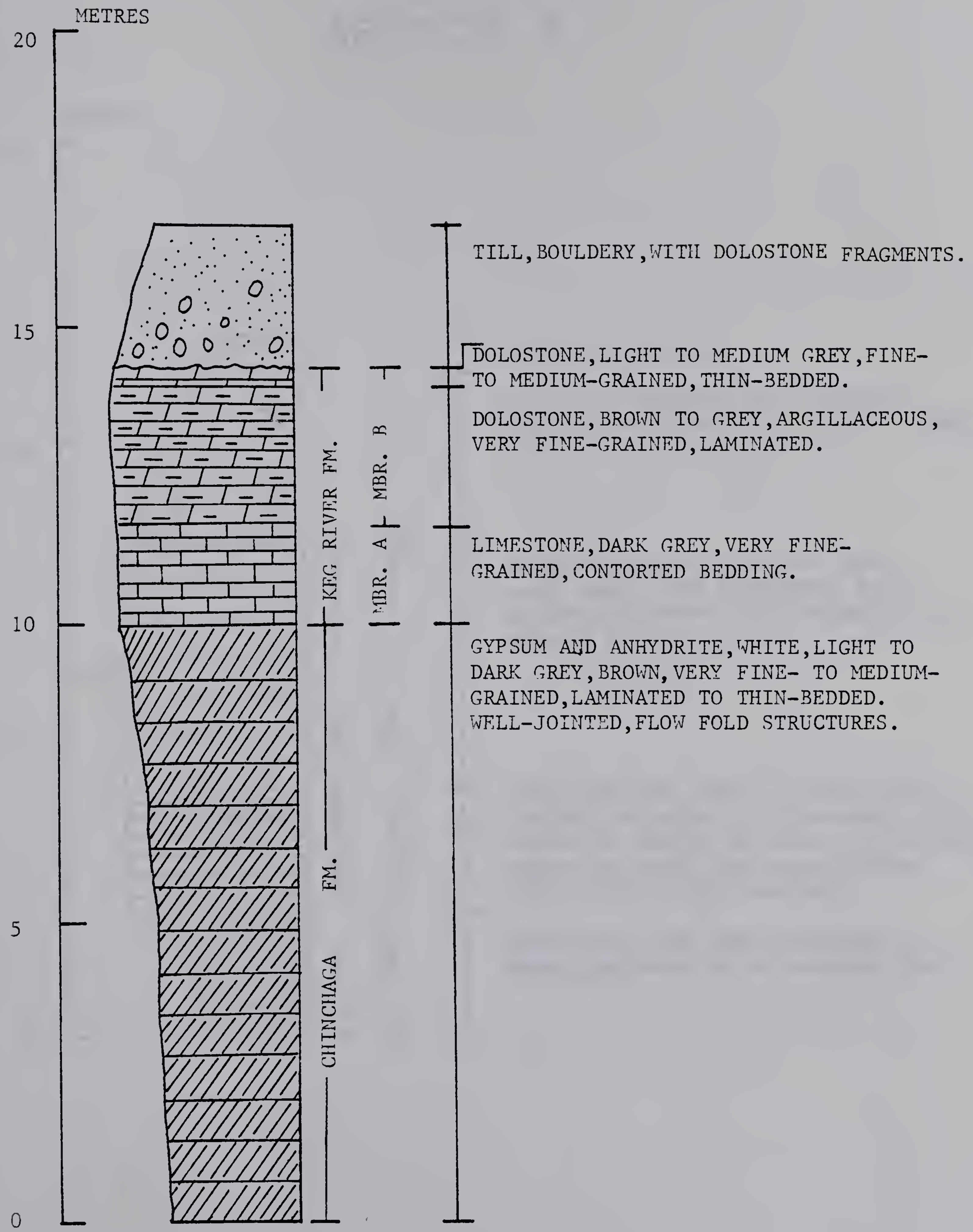


TILL, BOULDERY

GYPSUM AND ANHYDRITE, WHITE, LIGHT TO DARK GREY, BROWN, VERY FINE-GRAINED, LAMINATED TO THIN-BEDDED; MASSIVE TO THICK-BEDDED FROM A DISTANCE, SOLUTIONAL FEATURES SUCH AS KARREN FEATURES ARE COMMON ON THE ROCK SURFACE.

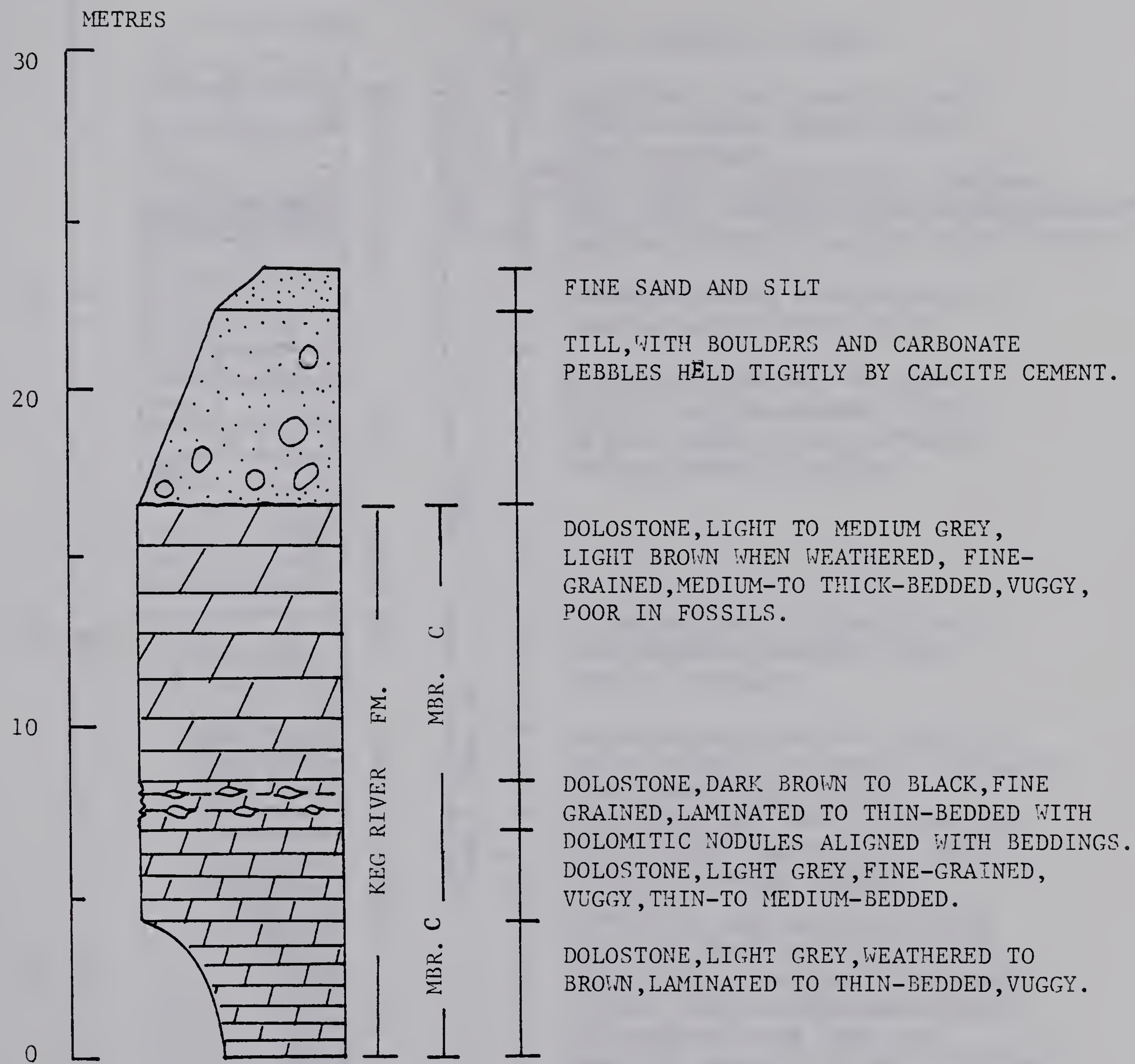
COVERED BY TALUS: APPEARED AS GYPSUM AND ANHYDRITE BEDS AS ABOVE.

SECTION 3

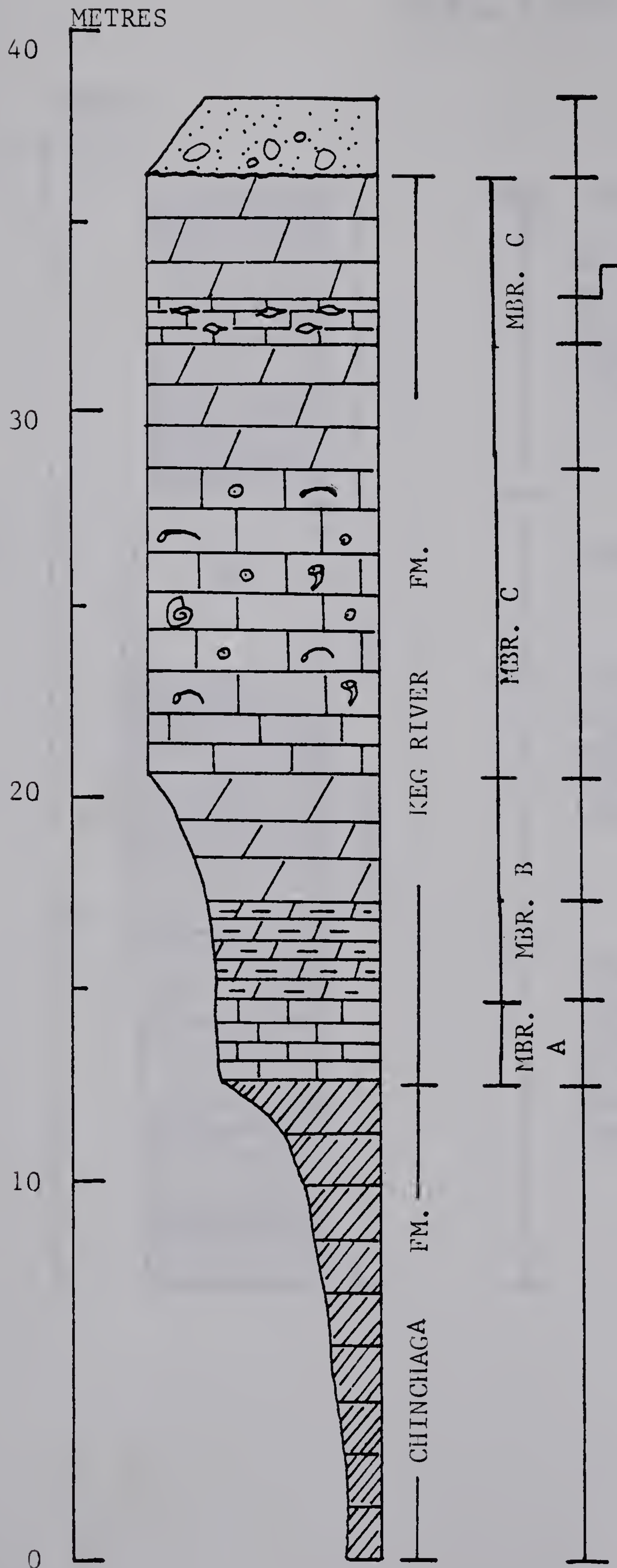


METRES
20
15
10
5
0
KEG RIVER FM.
MBR. A
MBR. B
CHINCHAGA FM.
TILL, BOULDERY, WITH DOLOSTONE FRAGMENTS.
DOLOSTONE, LIGHT TO MEDIUM GREY, FINE- TO MEDIUM-GRAINED, THIN-BEDDED.
DOLOSTONE, BROWN TO GREY, ARGILLACEOUS, VERY FINE-GRAINED, LAMINATED.
LIMESTONE, DARK GREY, VERY FINE-GRAINED, CONTORTED BEDDING.
GYPSUM AND ANHYDRITE, WHITE, LIGHT TO DARK GREY, BROWN, VERY FINE- TO MEDIUM-GRAINED, LAMINATED TO THIN-BEDDED. WELL-JOINTED, FLOW FOLD STRUCTURES.

SECTION 4


METRES
30
20
10
0
FINE SAND AND SILT
TILL, WITH BOULDERS AND CARBONATE PEBBLES HELD TIGHTLY BY CALCITE CEMENT.
DOLOSTONE, LIGHT TO MEDIUM GREY, LIGHT BROWN WHEN WEATHERED, FINE-GRAINED, MEDIUM-TO THICK-BEDDED, VUGGY, POOR IN FOSSILS.
DOLOSTONE, DARK BROWN TO BLACK, FINE GRAINED, LAMINATED TO THIN-BEDDED WITH DOLOMITIC NODULES ALIGNED WITH BEDDINGS.
DOLOSTONE, LIGHT GREY, FINE-GRAINED, VUGGY, THIN-TO MEDIUM-BEDDED.
DOLOSTONE, LIGHT GREY, WEATHERED TO BROWN, LAMINATED TO THIN-BEDDED, VUGGY.
KEG RIVER FM.
MBR. C
MBR. C

SECTION 5



TILL, BOULDERY, PEBBLY.

DOLOSTONE, LIGHT BROWN TO GREY, FINE-TO MEDIUM-GRAINED, VUGGY, VERY RESISTANT.

LIMESTONE TO DOLOMITIZED LIMESTONE, DARK GREY, LAMINATED, FINE-GRAINED, DOLOMITE OR CARBONATE NODULES ALIGNED WITH BEDDINGS.

DOLOSTONE, BROWN TO GREY, FINE-TO COARSE-GRAINED, THICK-BEDDED.

LIMESTONE, LIGHT GREY TO BROWN, FINE-TO MEDIUM-GRAINED, MEDIUM-TO THICK-BEDDED, FOSSILIFEROUS: BIVALVES, CORALS, CRINOIDS.

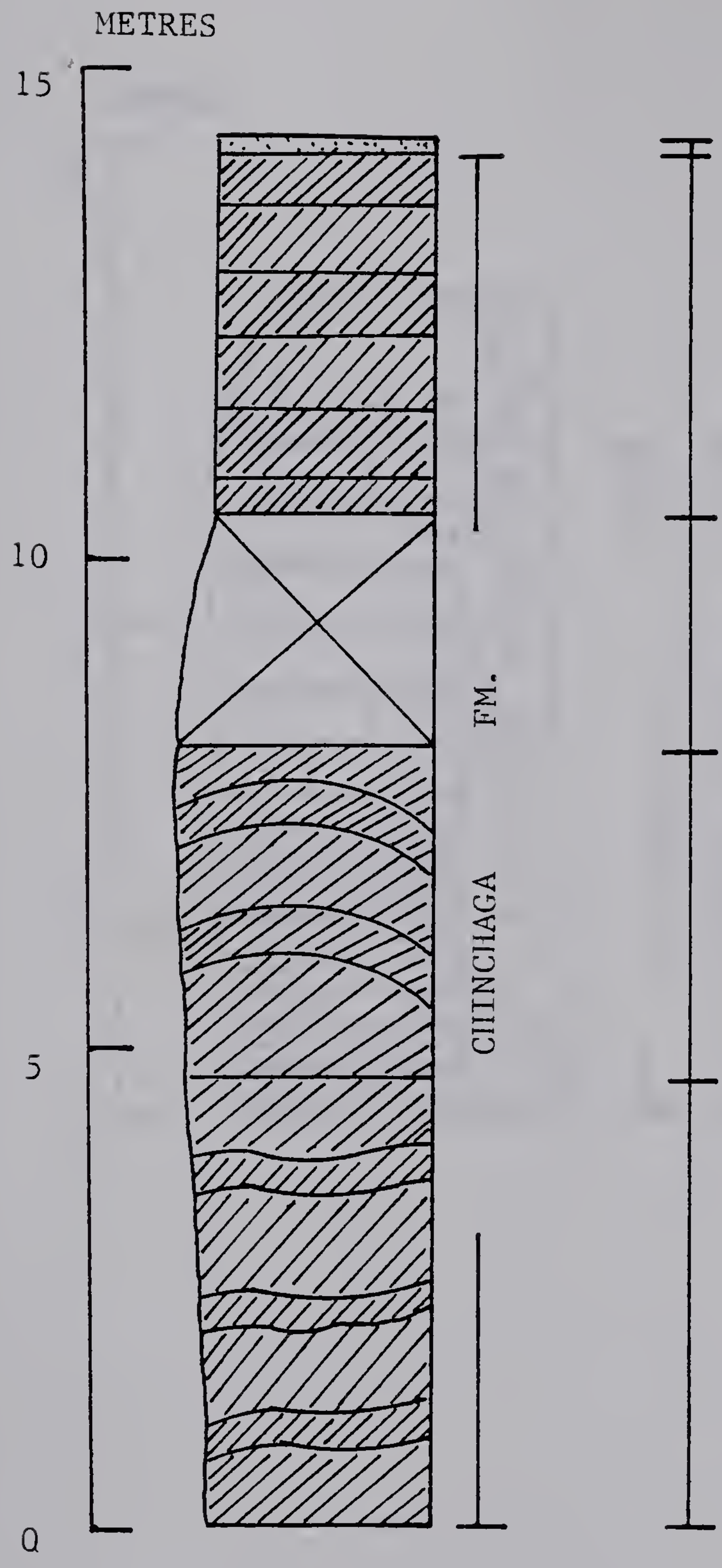
DOLOSTONE, BROWN TO GREY, FINE-TO COARSE-GRAINED, MASSIVE, VUGGY, CRINOID OSSICLES.

DOLOSTONE, ARGILLACEOUS, LIGHT TO MEDIUM BROWN, FINE-GRAINED, LAMINATED, INTERBEDDED CLAY LAYERS.

LIMESTONE, DARK BROWN TO BLACK, VERY FINE-GRAINED, LAMINATED TO THIN-BEDDED, CONTORTED BEDDINGS.

GYPSUM AND ANHYDRITE WITH FINE DOLOMITE GRAINS, WHITE, LIGHT TO MEDIUM GREY, LIGHT TO BROWN, COARSE-TO VERY FINE-GRAINED, LAMINATED TO THICK-BEDDED, SOME BEDS SHOW CONVOLUTE FOLDS, DISTINCT SOLUTIONAL FEATURES SUCH AS CAVES.

SECTION 6



OVERBURDEN

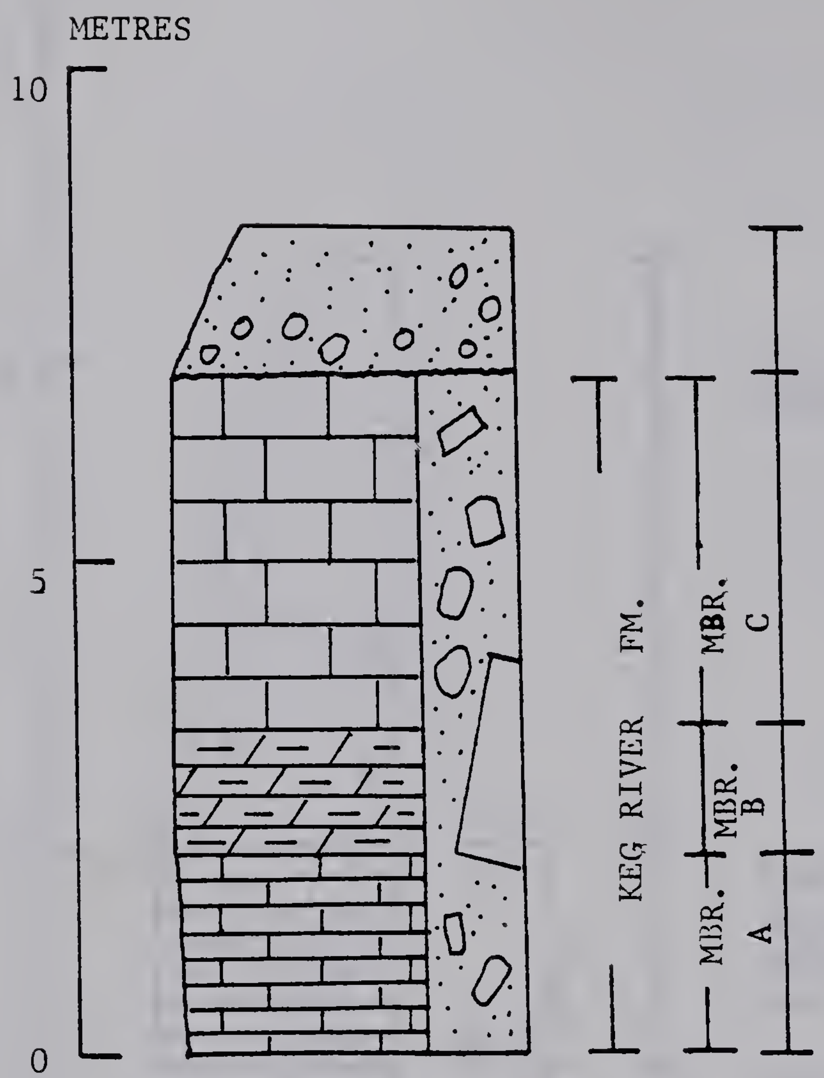
ANHYDRITE AND GYPSUM, PINKISH WHITE TO BROWN, VERY FINE-GRAINED, MEDIUM- TO THICK-BEDDED, LAMINATED SATIN SPAR LAYERS WITH CRYSTALS PERPENDICULAR TO BEDDING PLANES.

COVERED.

GYPSUM AND ANHYDRITE, WHITE TO LIGHT BLUISH GREY, FINE-TO COARSE-GRAINED, MEDIUM-TO THICK-BEDDED, GENTLY FOLDED, THIN-BEDDED SATIN SPAR CRYSTALS PERPENDICULAR TO BEDDINGS.

GYPSUM AND ANHYDRITE, POWDERY WHITE, INTERBEDDED WITH BROWN COLORED CLAY LAYERS, LAMINATED TO THIN-BEDDED, FINE- TO MEDIUM-GRAINED, NODULAR GYPSUM, SATIN SPAR LAYERS, PRESENCE OF FLOW FOLDS. SPRINGS RESURGE AT THE BASE.

SECTION 7



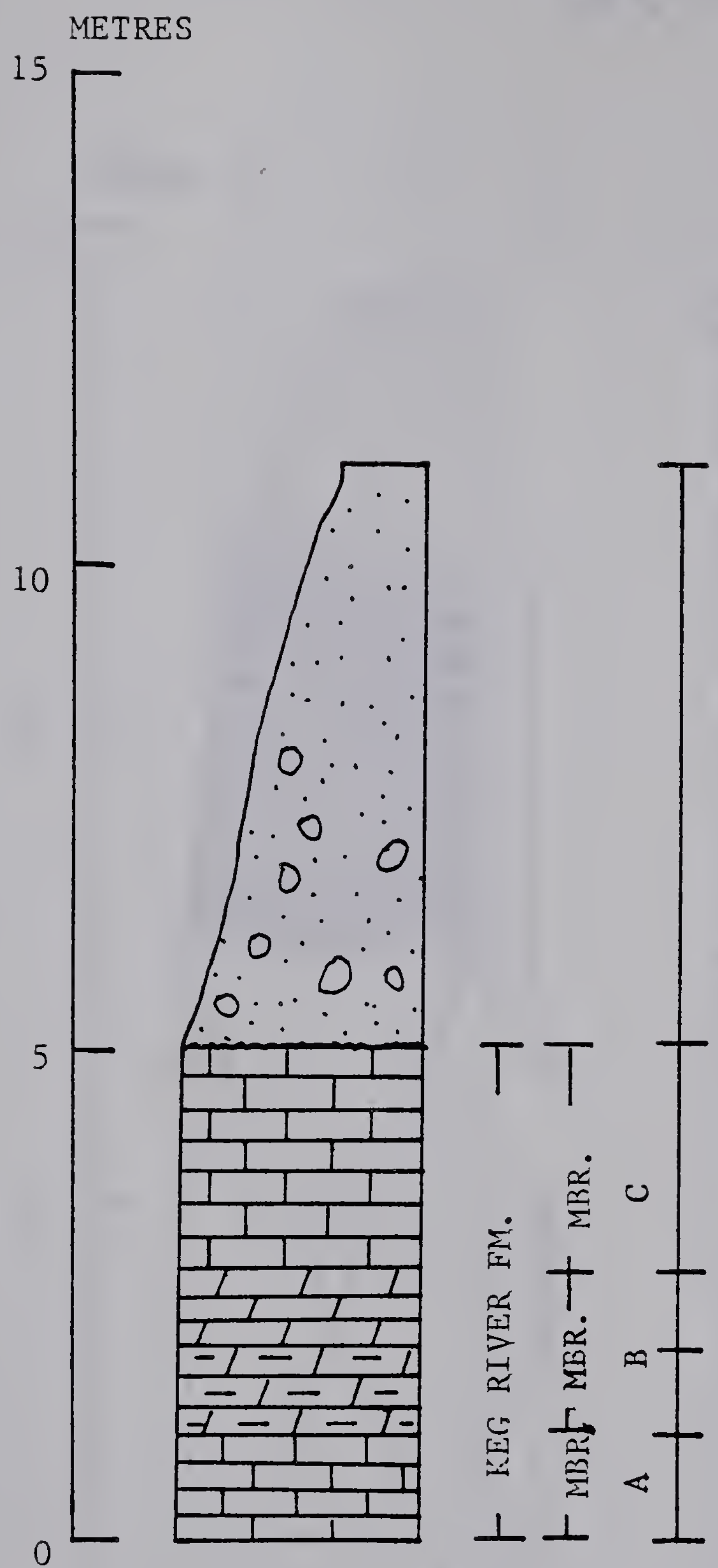
TILL WITH COBBLES UP TO 0.8M^3.

LIMESTONE, GREY, WEATHERED TO ORANGE BROWN, FINE-TO COARSE-GRAINED, THIN-TO MEDIUM-BEDDED. FOSSILIFEROUS. GENTLY FOLDED. CEMENTED BRECCIA TROUGH EXTENDED FROM THE TOP TO BOTTOM OF THE SECTION.

DOLOSTONE, LIGHT BROWN TO LIGHT GREY, ARGILLACEOUS, FINE-GRAINED.

LIMESTONE, DARK GREY, FINE-TO MEDIUM-GRAINED, LAMINATED, CONTORTED BEDDING, BRITTLE AND WEATHERED TO THIN FISSILE PLATES.

SECTION 8



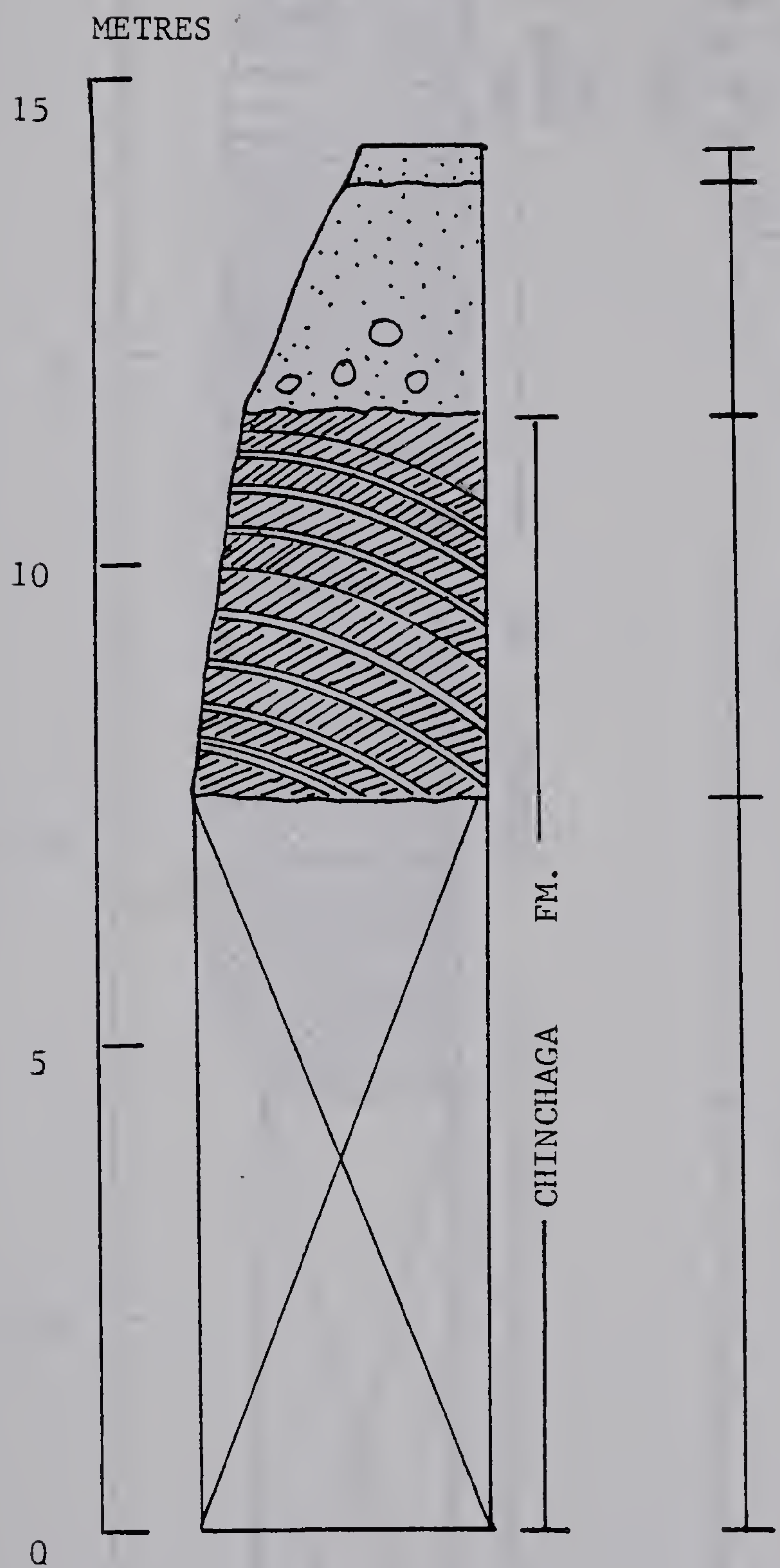
TILL WITH BOULDERS, CARBONATE PEBBLES.

LIMESTONE, MEDIUM GREY, FINE-TO MEDIUM-GRAINED, THIN-TO MEDIUM-BEDDED. FOSSILIFEROUS: TRILOBITES, CUPCORALS, BIVALVES, CRINOIDS.

DOLOSTONE, BROWNISH GREY, FINE-TO MEDIUM-GRAINED, THIN-BEDDED, VUGGY.
DOLOSTONE, BROWNISH GREY, ARGILLACEOUS, VERY FINE-GRAINED, LAMINATED.

LIMESTONE, DARK GREY, VERY FINE-GRAINED, LAMINATED TO THIN-BEDDED, CONTORTED BEDDING.

SECTION 9



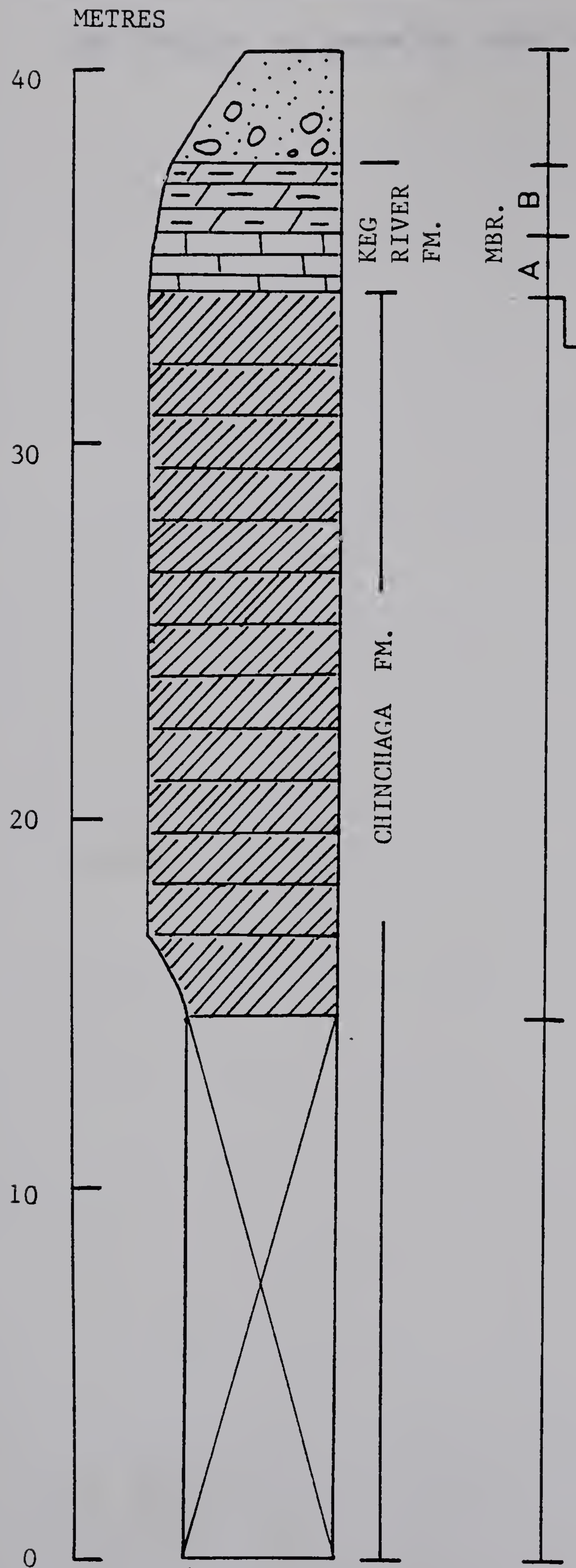
SAND

TILL, BOULDERY

GYPSUM, POWDERY WHITE, GREY, COARSE-TO FINE-GRAINED, INTERBEDDED WITH LAMINATED TO THIN-BEDDED CLAY, SELENITE AND SATIN SPAR CRYSTALS ARE COMMON, TIGHT CLOSED FOLD WITH FLOW FOLD FEATURES.

COVERED: APPEARS TO BE GYPSUM BEDS AS ABOVE.

SECTION 10



TILL, WITH GRANITIC BOULDERS.

DOLOSTONE, LIGHT BROWN TO GREY, ARGILLACEOUS, LAMINATED TO THIN-BEDDED.

LIMESTONE, DARK GRAY, VERY FINE-GRAINED, LAMINATED TO THIN-BEDDED, CONTORTED BEDDINGS AND BRECIATION, WEATHERED TO LIGHT GREY.

GYPSUM AND ANHYDRITE, WHITE, LIGHT BROWN, LIGHT TO DARK GREY, FINE-GRAINED, LAMINATED TO THIN-BEDDED, PRESENCE OF GRANULAR, NODULAR TO FINE-GRAINED DOLOMITE, HIGHLY JOINTED, SATIN SPAR CRYSTALS ARE PERPENDICULAR TO THE BEDDING PLANES, WEATHERED TO THIN FISSILE BEDS. SPRINGS APPEAR AT THE BASE.

COVERED, APPEARS TO BE GYPSUM AND ANHYDRITE AS ABOVE.

Appendix 3 : Computation of the maximum time required for the faults to develop near south Salt River Bridge area

Appendix 3

Volume of gypsum dissolved away beneath the triangular faulted block

= area times vertical displacement

= (16368 ft x 27456 ft) / 2 x 40 ft

= (4989 m x 8369 m) / 2 x 12.2 m

= 2.54×10^8 m^3

Specific gravity of the block (gypsum & anhydrite)

= 2600 kg/m^3

Total mass of gypsum & anhydrite dissolved away is

= 6.6×10^{11} kg

According to the hydrologic equation (Mathewson, 1981, p. 117):

Precipitation (P) = Runoff (Q) + Evapotranspiration (ET) + Infiltration (I) + Storage (S)

In order to obtain a maximum infiltration value, let the runoff (C) and storage(S) in this karst area be equal to zero.

Annual precipitation = 315 mm (Fort Smith Flight Service Station, 1981)

Annual evapotranspiration = 250 mm (Slivitzky, 1978, p. 25)

Thus, I = P - ET

= 315 mm - 250 mm

= 0.065 m / year

Total volume of water infiltrated into the triangular faulted block per year:

= 0.065 m / year x (4989 m x 8369 m) / 2

= 1.357×10^6 m^3 / year

Appendix 3 (cont'd)

Assuming the infiltrated water would be gypsum saturated before it discharged along the escarpment. Thus, the concentration of the water is

= 2070 mg / litre (Freeze & Cherry, 1979, pp. 280 - 284, Table 7-7)

= 2.07 kg / m^3

Total mass of gypsum & anhydrite loss per year

= 1.357×10^6 m^3 / year x 2.07 kg / m^3

= 2.81×10^6 kg / year

Total time span required for the fault to occur

$$= \frac{6.6 \times 10^{11} \text{ kg}}{2.81 \times 10^{6} \text{ kg / year}}$$

= 234980 years which is at least 124980 years before the beginning of the last glaciation (Sugden & John, 1976, p. 131)

Assuming the faulting is due to solution subsidence; thus, the average displacement per year

$$= \frac{12.2 \text{ m}}{234980 \text{ years}}$$

= 0.000052 m

= 0.052 mm / year

Appendix 4 : Computation of the minimum allowable roof span for caverns

Appendix 4

According to Davies (1951) :

$$L = (4FT / 3W)^{\frac{1}{2}}$$

L = roof span (unknown)

F = tensile strength (assuming stress that caused failure is the weight of the span itself)

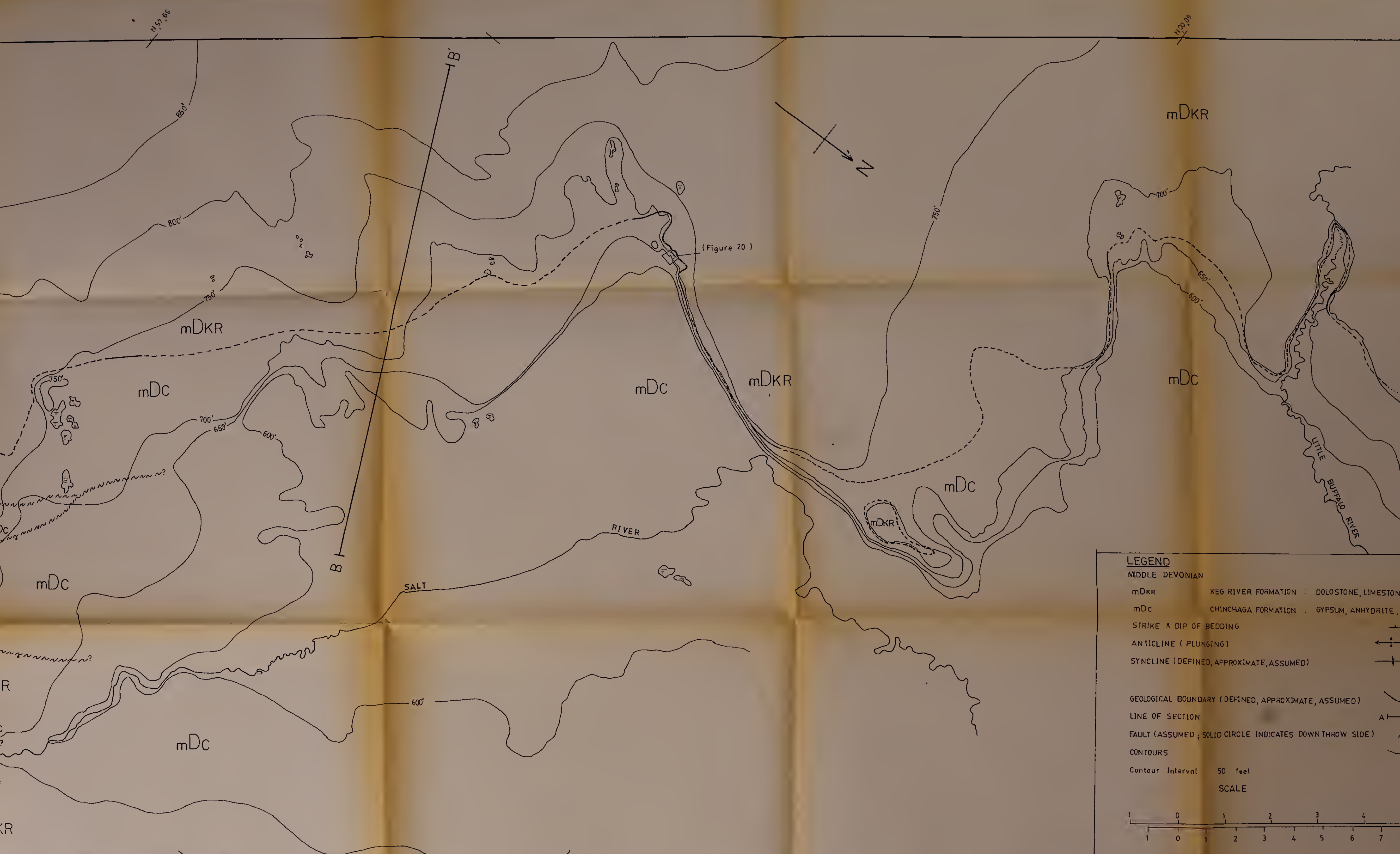
T = thickness of span (assuming it has detached from the overlying strata)

W = unit weight of the material forming the span

Rock properties of the dolostone in the Fire Tower Collapse Doline:

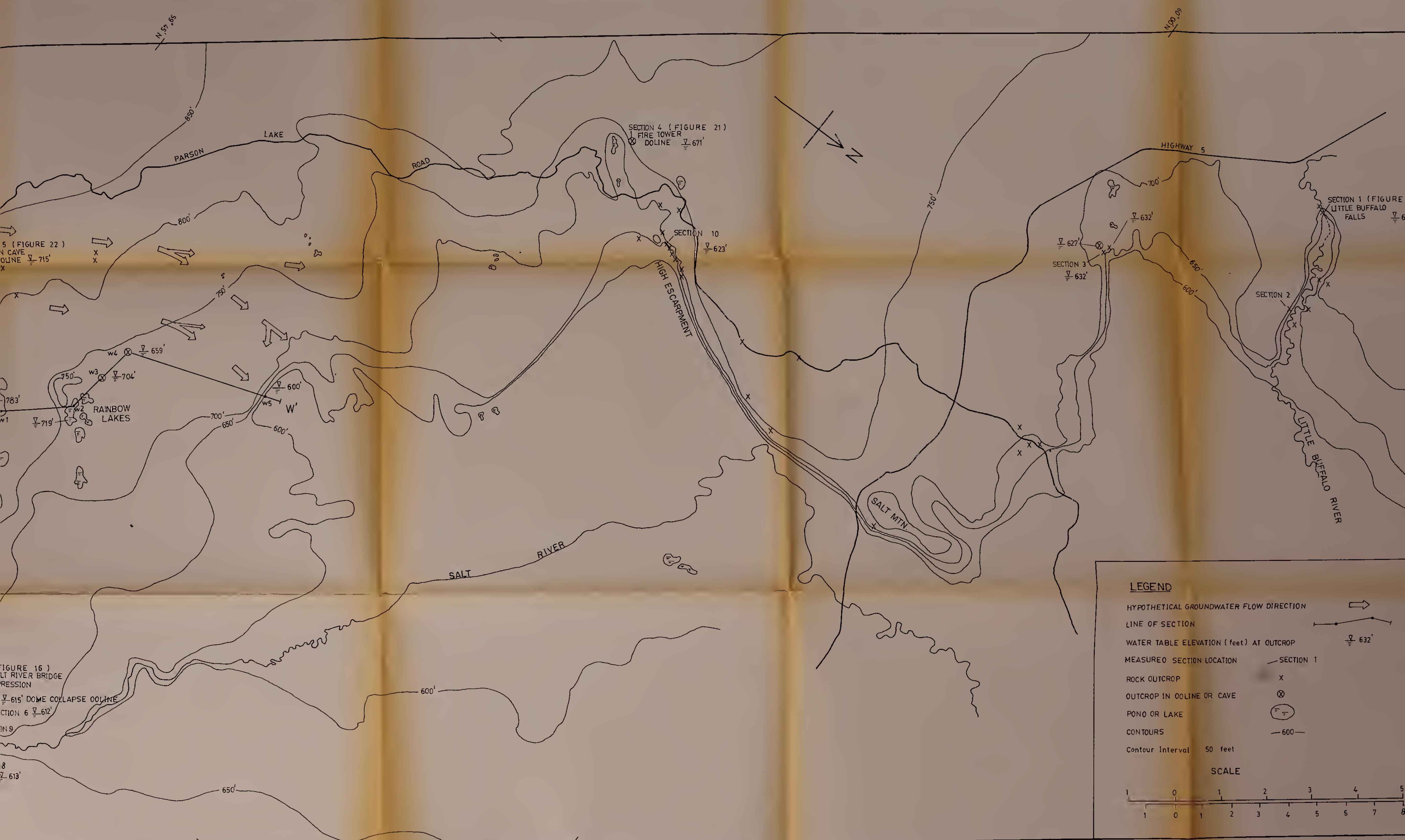
1. Average unconfined compressive strength = 107.6 MPa (Cruden, 1980)
2. According to Griffith Brittle Failure Criterion (Roberts, 1977, p.52), the unconfined tensile strength is about 1/8 of the unconfined compressive strength; thus, F = 13.45 MPa
3. Unit weight, W = 24.53 KN/m^3 (Cruden,1980)
4. Maximum thickness of bed, T = 0.9 m (Appendix 2, Section 4)

Substitution of the above rock properties of the dolostone into Davies' equation, a roof span of 26 m wide is obtained.



URE 6 : GEOLOGIC MAP OF THE STUDY AREA, WOOD BUFFALO NATIONAL PARK

(P. C. TSUI



(P. C. TSUI,

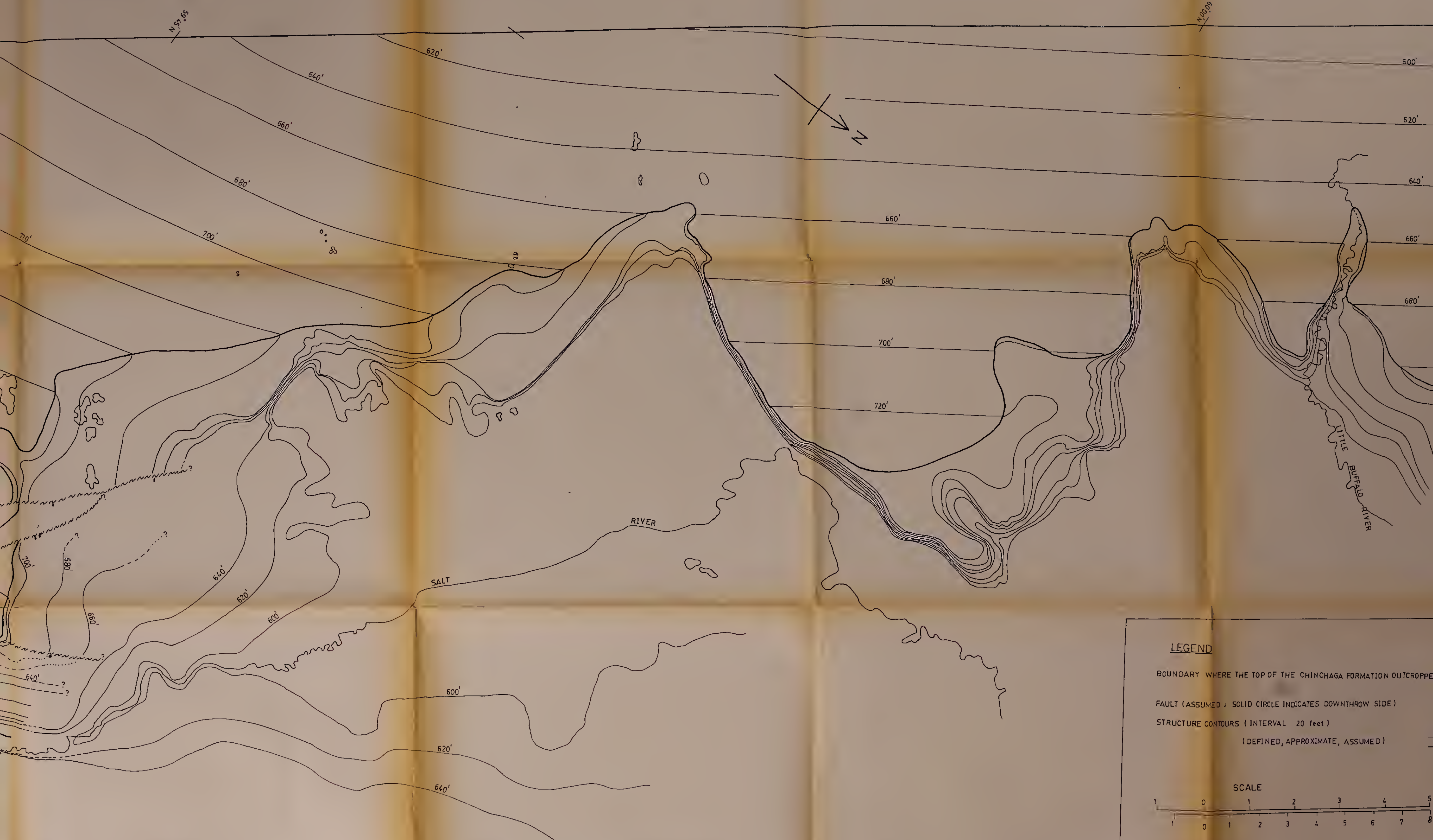
OCATIONS OF OUTCROPS AND MEASURED SECTIONS AND THEIR CORRESPONDING WATER TABLE

Figu

: GEOMORPHOLOGICAL MAP OF THE STUDY AREA, WOOD BUFFALO NATIONAL PARK

(P. C. TSUI, 198

Figure 9



GURE 10 : STRUCTURE CONTOUR MAP ON TOP OF CHINCHAGA FORMATION

(P. C. TSUI,

Figure